M.I. Budyko

GLOBAL ECOLOGY

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Preface

Following the rapid development of anthropogenic changes in the environment in recent years there emerged a heightened interest in problems of global ecology. This is why it was important to write a book devoted entirely to the problems of that science, even though this has presented some difficulties.

Global ecology as a distinct scientific discipline is still in a state of development, and its boundaries are not yet clearly defined. While not seeking to encompass all branches of that science, the author has found it appropriate to emphasize the central problem of world ecology, namely, the circulation of energy and of different types of matter within the biosphere. Wide use has been made of the author's own earlier studies, including relevant sections of *Thermal Balance of the Earth's Surface* (1956), *Climate and Life* (1971), and *Climatic Changes* (1974).

These studies are needed to solve current problems of global ecology, as well as to explain the patterns of the biosphere's evolution in the past and to forecast its possible changes in the future.

Particular emphasis has been given to the geophysical aspects of global ecology, since the author possessed more data on that subject than on other aspects of that science.

I

The Biosphere

1. The Problems and Tasks of Global Ecology

Ecology is the science that studies mutual relations between organisms and their environment. The word is derived from the Greek *oikos*, which means household or dwelling. While the term "ecology" was proposed as early as 1869 by Ernst Haeckel, a prominent biologist, it is only in the beginning of this century that systematic studies of interactions between organisms and their environment were carried out.

During the first half of the twentieth century problems relating to the interactions of living organisms with their environment were studied by specialists in such sciences as botany, zoology, soil science, geography, paleontology, and geochemistry. In many cases the authors of these studies did not view them as belonging to ecology.

The position of ecology has been altered during the second half of this century, when the influence of human activities on the environment greatly increased. The pollution of the atmosphere, of oceans and of inland water bodies, the destruction of the Earth's natural plant cover and the disappearance of many types of animal, the soil erosion and other consequences of man's economic activity have all drawn attention to the problems of the environment and have contributed to a substantial growth of ecological studies. The major problems of contemporary ecology include the study of anthropogenic changes in man's environment as well as the scientific substantiation of methods for preserving and improving that environment in the interests of mankind.

In order to solve both these and other problems a fundamental difficulty must be overcome that is absent in the case of many other scientific disciplines. The study of the interaction between living organisms (including man) and their environment requires the use of materials and methods drawn from many other sciences including biology, geography, geophysics, geochemistry, geology, economics, and engineering.

Today when a vast increase in the volume of information in each scientific field produces a further subdivision of these fields into a growing number of distinct sciences, the possibilities for a synthesis of materials drawn from many different fields of knowledge in order to solve particular problems are becoming more limited. Yet without such a synthesis further progress in modern ecology is not possible.

One of the ways in which this difficulty may be overcome is to carry out complex studies of ecological problems by teams including representatives of different sciences. This is by no means a simple matter, since its success largely depends on a close integration of individual contributions and requires a high degree of organization.

In recent years broad studies of complex ecological problems that will have a large influence on that science's further development have been undertaken in a number of countries.

The major general concept of ecology is that of an ecological system (an ecosystem) that includes a group of mutually interdependent living organisms together with those elements of the environment whose influence on them is particularly strong and that themselves also depend in some measure on the activities of such organisms. In many studies the term "ecological system" is replaced by another closely related term, namely "biogeocenosis" (or else "geobiocenosis").

The boundaries and content of ecological systems that are accepted in ecological studies should be recognized as purely conventional. For each group of living organisms a multiplicity of relations exists with other organisms and elements of the environment from which it is possible to select either a wider or else a more limited set of interrelations when identifying an ecological system. This will determine the scale of the ecological system and the number of components that it includes. At the same time it should be kept in mind that in certain cases ecological systems possess more or less well-defined natural boundaries that are clearly expressed, as for example in the case of small closed bodies of water or islands with a relatively homogeneous surface. In such cases the problem of defining the boundaries of an ecological system and of its components becomes simpler. Depending on the objectives that researchers pursue ecological systems of various sizes may be considered, including the system that covers the entire sphere of existence of living organisms, i. e., the biosphere, consisting of the atmosphere, hydrosphere, and lithosphere within which either living organisms or the products of their activities are found. Compared with the dimensions of the globe itself it represents a thin envelope whose thickness does not exceed several tens of kilometers.

Global Problems of Ecology. The overwhelming majority of ecological studies are concerned with local problems of ecology, i.e., with studies of interaction between organisms and the environment within regional ecological systems. A relatively small number of researchers have been engaged in studies of global problems of ecology relating to the biosphere as a whole or else to its major parts.

An interest in global ecological problems has emerged in recent years when it became apparent that man's economic activities have begun to influence large-scale natural processes. Even though natural conditions have been altered over a major part of the Earth's surface quite some time ago as a result of anthropogenic influences these were, until recently, largely aggregates of local changes that gradually spread over large areas following the extension of the sphere of man's economic activities. For example, the destruction of forests on one continent did not influence the forests of other continents, and the construction of dams on individual rivers did not influence the run-off of other rivers not connected with them.

A different situation arises when man begins to influence nature's global processes. Influences of the environment of one region may change natural conditions in other regions at great distances from the place of man's activities.

Some examples of man's influence on large-scale natural processes were known earlier. In particular the destruction of migratory birds in individual countries at middle latitudes has altered the fauna of the tropical countries that were visited by such birds. Similar situations existed in the case of certain migratory sea animals and fish. Nevertheless these examples relate to those elements of the biosphere whose influence on the biosphere as a whole is comparatively modest.

In recent years it has been established that as a result of human activities the chemical composition and physical state of the atmosphere and of the oceans have begun to change. This raises the possibility of anthropogenic changes in other components of the biosphere. In such cases, moreover, an intensive horizontal mixing of air and of upper layers of oceanic waters may cause anthropogenic influences on the atmosphere and the ocean to spread to very large distances from the localities where they occur.

Since atmospheric processes exert a substantial influence on all components of the biosphere, including living organisms, it is evident that large-scale changes in the atmosphere-ocean system will inevitably lead to changes in the biosphere as a whole that may prove to be unfavourable and even catastrophic for humanity.

Since unregulated human influences on large-scale processes within the atmosphere and the ocean may ultimately lead to a global ecological crisis there is an evident need for studies of the possibility of large-scale anthropogenic changes of the biosphere in order to make quantitative forecasts of these changes under different alternatives of economic development. Such forecasts are needed to select optimal paths of technological progress whose influence on the environment will not be destructive.

The problems of global ecology are not limited to the influences of human activities on the biosphere.

Methods of Investigation. In the geological past important changes occurred in the physical state and chemical composition of the atmosphere, ocean, and upper layers of the lithosphere that were partly caused by biological processes, and they themselves influenced living organisms. Studies of global ecology must therefore also include a clarification of the patterns of the evolution of the biosphere.

In studying the current state of the biosphere wide use is made of data collected both by stationary observation posts and by expeditions.

Information concerning physical processes taking place in the atmosphere is obtained through meteorological observations that cover all regions of the globe. They are effected through a worldwide network of permanently operating meteorological stations located on all continents and on many islands. These include more than 10,000 climatic stations and several tens of thousands of meteorological posts. Daily observations of air temperature and humidity as well as of atmospheric precipitation, clouds, winds, pressure, and other meteorological elements are made at the climatic stations. More limited programmes of observations are carried out at meteorological posts. Frequently they are limited to observations of atmospheric precipitation.

A global network of actinometric stations forms an important part of the system of meteorological observations. This includes approximately one thousand stations at which observations of solar radiation and its transformation are carried out. Similarly, aerological stations play a substantial role in studies of the atmosphere. There is approximately one thousand such stations which employ radiosondes and other instruments in order to observe meteorological elements within the free atmosphere up to heights of 30-40 kilometers.

A number of stations carry out observations of the ozone in the atmosphere, of atmospheric electricity, and of the air chemical composition.

Data collected by land meteorological stations are supplemented by observations made on the oceans, where weather ships operate continuously in various regions of the World Ocean. The number of weather ships is not large, and an important role in studying the atmosphere over oceans is also played by meteorological information received from other ships at sea.

Data pertaining to the physical state of ocean waters are largely provided by research expeditions through measurements taken from board the ships of temperature, salinity, optical properties, and other hydrological elements at various depths. A complex of such measurements is usually called an oceanographic "station".

There have been approximately 200,000 such oceanographic stations carried out since studies of the ocean have begun. A similar number of stations have been carried out in seas, where their distribution is rather uneven. While observations in the oceans and seas provide a wide range of information on the physical state of waters, its volume is far less than that produced by observations carried out in the atmosphere. As a result less knowledge has been obtained of physical processes in the oceans, in spite of the fact that they are more stable than those that occur in the atmosphere. Information on the state of oceanic waters at depths of more than two kilometers is especially scarce, since most studies are carried out at smaller depths.

The principal characteristic of the hydrological regime of land is the volume of river run-off. The corresponding measurements are carried out at hydrological stations and observation posts of which there are approximately 60,000 at the present time. The hydrological observation network is very uneven. While in Europe and North America there is one observation post per 1-2,000 square kilometers, in many tropical countries the corresponding ratio is 1 per 10,000 square kilometers or more.

Far fewer observations are made of the biosphere's biological components, although there has been substantial progress in this respect over recent years as a result of the implementation of the International Biological Programme. One of its objectives was to obtain data on the productivity of living organisms in different geographical regions.

The biosphere's biological components have experienced far greater changes as a result of human activities than its nonbiological components, and data on the flora and fauna as they exist in natural conditions over major parts of the continents' surface can be obtained primarily in nature preserves whose number is limited.

In some of the nature preserves and other sparsely settled regions there are stationary observation centres engaged in systematic observations of the biomass of various plants and animals which make it possible to estimate productivity of organisms. While data collected there are supplemented by data provided by research expeditions, available information on biomass and productivity is often insufficient and does not permit generalizations. In particular more or less comprehensive and reliable data on the productivity of the natural plant cover are available only for several tens of regions on the surface of the continents. Information concerning biological processes in the ocean and especially in regions that are seldom visited by research ships is even less adequate.

In recent years artificial Earth satellites have been employed in carrying out observations of the state of the biosphere. Special-purpose equipment has made it possible to receive global information concerning the atmospheric meteorological regime, including data on clouds, the radiation balance, air temperature and humidity. At the same time data are obtained on the temperature of the ocean's surface as well as its waves. Satellites also provide information on the hydrological regime of land and the state of the plant cover as well as on other biospheric phenomena.

As methods employed by such satellites are improved their significance for studying the biosphere will continue to grow.

In addition to data on the biosphere's current state, research on problems of global ecology requires empirical data on changes in the biosphere that have taken place in the past. Such data may be provided by studies of paleogeography, historical geology, geochemistry, and paleontology, in which much emphasis is given to the study of sedimentary deposits. While information on the biosphere in the distant past is quite substantial, it is very uneven. As a result many important patterns of natural conditions of past ages are still unfathomed.

Generalizations of empirical data on large-scale processes in the biosphere have made it possible to study patterns of transformations of solar energy on the Earth's surface, in the atmosphere and hydrosphere, to determine the components of the water exchange for the Earth as a whole as well as for its individual regions, and to estimate the balance of processes bearing on the creation and destruction of organic matter. The results of these studies are extremely important in clarifying the mechanism that determines the biosphere's structure and its evolution.

Aside from generalizations of empirical data studies of large-scale biospheric processes have been carried out through the numerical modelling method in recent years.

Important successes have been achieved in developing numerical models of processes taking place in the atmosphere in connection with the development of the theory of climate. The more sophisticated models represent physical processes taking place in the oceans. There are models that represent not only elements of the meteorological regime but also spatial and temporal distributions of elements of the hydrological regime on land, including the volume of river run-off.

It may be expected that in the very near future a model of the theory of climate will be constructed that will cover all major physical processes developing in the atmosphere and hydrosphere. This will greatly expand possibilities for studying casual linkages between abiotic components of the biosphere.

The problem of constructing numerical models of biological processes is more complicated than modelling of physical processes within the biosphere.

In recent years numerical models have been developed of photosynthesis within the layer of the Earth's plant cover which make it possible to calculate the productivity of autotrophic plants. The theory of the dynamics of animal populations is developing rapidly which attributes oscillations in their numbers to the influence of various external factors. Studies have been initiated in the numerical modelling of ecological systems of various complexity.

Existing numerical models of biological processes are employed primarily to study local ecological problems. Their use for studying large-scale processes in the biosphere is only beginning.

At the present time global ecology is in its initial stage, and an historical survey of its development is therefore a task for the future. One should nevertheless mention studies that have made possible the emergence of that science, of which a substantial number have been carried out by Russian and Soviet scientists.

The first known studies of interaction between living organisms and their environment in different natural zones were carried out by Vassily Vassilyevich Dokuchayev. In a series of lectures published in 1898-1900 and reprinted in 1948 in the form of a collection, he considered a system of natural zones that correspond to different types of the soil-formation process. Dokuchayev established relations between natural zones and climatic factors, and constructed for each natural zone a table describing soils, vegetation, fauna, and other components of the biosphere.

The works of Vladimir Ivanovich Vernadsky (1926, etc.) laid the foundations for the theory of the biosphere seen as an integral system whose development is largely determined by the activities of the living organisms. It was Vernadsky's view that following the appearance of man, the biosphere has been gradually shifting to a new state as a result of human activities which he called the Noosphere, i.e., the sphere of reason.

It was only in the second half of our century that the great significance of Dokuchayev's and Vernadsky's research for the development of ecology and in particular for studying of global problems of ecology was fully understood.

Dokuchayev's studies were continued by G. F. Morozov whose principal work, entitled *Knowledge About Forests* (1912), created the foundations of forest ecology. Subsequently the studies of G. N. Vysotsky added greatly to forest ecology.

Other geobotanic studies of importance for the development of global ecology include the works of V. N. Sukachyov (1945, 1964, etc.) concerned with biogeocenoses, those of Ye. M. Lavrenko (1949, 1971, etc.) which examine the proerties of the phytogeosphere (the sphere of existence of plants), and those of N. I. Bazilevich and L. Ye. Rodin (1967, etc.) that presented detailed data on the biomass and productivity of different types of plant cover.

In addition to geobotanic studies, studies of the ecology of animals contributed greatly to knowledge of global ecology. These were initiated by C. Elton (1927, etc.) and include the monographic studies of D. N. Kashkarov (1938), S. A. Severtsev (1941), D. Laek (1954), and G. V. Nikolsky (1974, etc.).

The ideas of Dokuchayev in the area of soil studies have been developed further in the works of I. P. Gerasimov (1945, 1960, etc.) which view the soil cover as an element of the biosphere that is closely related to its other components. This line of research also includes the works of V. R. Volobuyev (1953, etc.) and V. A. Kovda (1973, etc.), which are concerned with the process of soil formation as influenced by external factors.

A comprehensive picture of interaction among natural processes that lead to the formation of geographical zones emerged in the studies of A. A. Grigoriyev (1966, etc.) which established the dependence of natural zones on such elements of climate as solar radiation and precipitation. Other studies concerned with geographical zones include those of V. B. Sochava (1974, etc.), in which the method of systems analysis has been employed for study of natural zones.

An overall survey of contemporary conceptions of the biosphere is provided in the books by S. V. Kalesnik (1970, etc.). A monographic study written under the supervision of I. P. Gerasimov (1971) contains imformation on the biosphere's resources on the territory of the USSR as well as an analysis of the problem of man's influence on the biosphere.

A promising idea of the application of the methods of biophysics to studies of the biosphere has been elaborated in the publications of G. F. Khilmi (1966, 1969, etc.).

The problem of patterns of the evolution of the biosphere has been studied in the works of A. P. Vinogradov (1967, etc.), K. K. Markov (1960, etc.), and A. I. Oparin (1957, etc.). S. S. Schwartz (1973) has studied the ecological mechanism of evolution of living organisms.

The vital problem of anthropogenic changes in the biosphere has attracted much attention among Soviet researchers. In particular that problem is considered in the monographs of D. L. Armand (1966), Ye. K. Fyodorov (1972), Yu. A. Izrael (1974), and of many other authors. Numerous monographs concerning this problem that were written in other countries include the works of P. Duvigneaud and M. Tanghe (1967), of B. Commoner (1971), and of K. Watt (1968). The important practical problem of designing an optimal system for monitoring the state of the environment was considered in the studies of Yu. A. Izrael (1974, etc.) and of I. P. Gerasimov (1974).

Studies of problems of global ecology require data on the transformation of solar energy in the Earth's biosphere and water balance. Information on the energy balance is contained in a number of monographs and atlases, including *The Atlas of the Earth's Heat Balance* (1963). Detailed data on the water balance of land and oceans are presented in the monograph by a group of authors entitled *The World Water Balance and Water Resources of the Earth* (1974) and in the studies of M. I. Lvovich (1974, etc.). Data on energy and water balances were employed in studying global ecological problems (Budyko, 1956, 1971, 1974; T. Odum, 1971).

Data needed to study global ecological problems are found in a number of generalizing studies in ecology, of which the most comprehensive is the monograph by E. Odum (1971).

Thus studies of global ecology combine data from geography, biology, and a number of other sciences.

It has already been noted that the principal task of global ecology is to forecast possible changes in the biosphere under the influence of human activities in various stages of economic development. Since long-term economic planning activities of large capital investments will be greatly influenced by such forecasts, the latter must be highly reliable and this requires a further development of global ecology, with particular emphasis on the following lines of research:

1. Studies of components of the biosphere in order to produce detailed quantitative descriptions of such components for all regions of the world.
2. Studies of energy cycles and of the circulation of major types of mineral and organic matter in the biosphere for different geographical regions.
3. The construction of numerical models for each component of the biosphere that help study processes taking place over large regions of the Earth. This should lead to the construction of a complex numerical model for the biosphere as a whole that takes into consideration the mutual interaction of all components of the biosphere.
4. The collection of empirical data describing the state of the biosphere in the geological past with a view to identifying patterns of its evolution.
5. The application of numerical models to calculate past changes in the biosphere in order to widen the knowledge of the mechanism of the biosphere's evolution. Also comparisons of the results of calculated changes in the biosphere with empirical data with a view to establishing the reliability of numerical models in forecasting future changes of the biosphere.
6. The application of numerical models to forecast anthropogenic changes in the biosphere as well as to forecast changes produced by natural causes in order to define optimal paths of economic development.
7. A search for methods of influencing large-scale processes within the biosphere in order to create a global system for regulating the biosphere in the interests of human society.

The problems listed are still far from being solved but the available data that already exist in relation to each makes it possible to tackle the central problem of global ecology. At the same time the nature of these data limits possibilities for arriving at reliable information concerning future changes in the biosphere. It is therefore necessary to continue such studies along each of the various lines of research.

2. The Structure of the Biosphere

The Atmosphere. The atmosphere surrounding the terrestrial globe largely consists of nitrogen and oxygen. It also includes small quantities of carbon dioxide and ozone as well as of other gases. The atmosphere exerts a great influence on the physical, chemical, and especially biological processes that take place on the Earth's surface and in the hydrosphere. The biological processes are especially sensitive to the following factors: the oxygen that is contained in the atmosphere and that is employed in the respiration of organisms and in the mineralization of organic matter; the carbon dioxide that has been employed in the photosynthesis of autotrophic plants; and the ozone which inhibits ultraviolet radiation from the Sun that is harmful to organisms. Without the atmosphere the existence of any complex living organisms would be impossible. This may be seen in the case of the Moon.

Researchers who study the Earth's past believe that initially the Earth, too, did not possess any atmosphere and that it was eventually created by the lithosphere in the course of our planet's evolution.

The development of the atmosphere is closely related to geological and geochemical processes, and to the activities of living organisms. While an important part of the gases that constitute today's atmosphere (nitrogen, carbon dioxide, and water vapour) arose in the course of volcanic activities which brought them from the depths of our planet, the atmospheric oxygen has evidently appeared at a later stage of the Earth's history as a result of the photosynthesis of plants.

In its turn the atmosphere has exerted a great influence on the evolution of the lithosphere: throughout its entire history it served as a powerful factor in the physical erosion by wind that has moulded the surface of the continents. Atmospheric precipitation and winds have contributed to that process by carrying small particles of rock over large distances. Changes in temperature and other atmospheric factors have also played an important role in eroding rocks.

Chemical erosion has played an even greater role in the evolution of the lithosphere. This is especially true of the influence on rocks of atmospheric oxygen and atmospheric water.

The evolution of the atmosphere has played a large role in determining the evolution of the hydrosphere, since the water balance of water bodies is influenced directly by precipitation and evaporation processes. On the other hand, atmospheric processes, too, have been influenced by the state of the hydrosphere, and especially by the state of the oceans. Generally, the evolution of the atmosphere and the hydrosphere represents a single process.

Atmospheric factors have played an important role in influencing the evolution of living organisms. More or less complex organisms could arise only after an amount of oxygen had accumulated in the atmosphere that was sufficient for the development of the aerobic processes needed for the energy absorption of complex organisms.

The activities of autotrophic plants have been closely associated with the carbon dioxide content in the atmosphere. Changes in that content in the course of the atmosphere's evolution have exerted a direct influence on the structure of the Earth's plant cover.

The existence of all living organisms has depended on heat and humidity regimes that change substantially with changes in climate. The important role that this plays in influencing the evolution of organisms is illustrated in Chapters VII and VIII.

The atmosphere's chemical composition is characterized both by permanent and variable components. The most important variable component is water vapour, most of which is concentrated in the troposphere. The average water vapour content in a vertical air column at moderate latitudes is approximately 1.6-1.7 centimeters of the "layer of precipitated water". Changes in the water vapour content over space and time are determined by interaction among evaporation, condensation, and horizontal movement.

Ozone exerts a definite influence on the atmospheric processes. It is largely concentrated in the stratosphere where it provides for the absorption of ultraviolet solar radiation. Another important variable component of the atmosphere is carbon dioxide, whose content depends on the life processes of plants and is also associated with its solubility in sea water (gas exchanges between the ocean and the atmosphere). Normally variations in its content are not large.

A marked influence on radiation processes in the atmosphere is exerted by atmospheric aerosols, i.e., particles that are suspended in the air, whose dimensions vary from tens of angströms to several tens of micrometers. Aerosols are found both in the troposphere and in upper layers of the atmosphere, and emerge as a result of the "pollution" of the atmosphere from the Earth's surface, including industrial waste and volcanic activities, and due to cosmic factors. Aerosol concentration diminishes rapidly with altitude, with numerous secondary maxima being imposed on this process due to the existence of aerozol layers.

The major part of the atmosphere is found within the troposphere, where temperature decreases with altitude. The altitude of the troposphere varies from 8-10 kilometers at high latitudes to 16-18 kilometers in the equatorial zone.

Physical processes within the troposphere determine changes in the weather and exert a major influence on the climatic conditions in different regions of our planet. These processes include the absorption of solar radiation; the formation of fluxes of long-wave radiation, which is dissipated into outer space (and which changes little in the higher layers of the air); and the water exchange that is associated with the formation of clouds and with precipitation. The factors that determine the genesis of climate are considered in Chapter III, while the patterns of the water exchange are considered in Chapter IV.

The Hydrosphere. Water in its various states is a major element of all components of the biosphere, and one of the leading factors in the existence of living organisms. The major part of the water of the biosphere is found in the World Ocean, which occupies approximately seventy-one per cent of the Earth's surface. Its average depth being 3.7 kilometers, the overall mass of ocean's water exceeds 1,300,000,000 cubic kilometers. A substantial quantity of water is contained in ice sheets—approximately

24,000,000 cubic kilometers, ninety per cent of that volume being found in Antarctica. Almost as much water is found under the ground at depths of up to two kilometers.

The remaining components of the world's supply of water are much smaller. The amount of water in lakes is approximately 0.18 million cubic kilometers, the volume of water in the atmosphere is 0.013 million cubic kilometers. In rivers, it is 0.002 million cubic kilometers, and the quantity of biological water (i.e., water contained in living organisms) is about 0.001 million cubic kilometers.

It is thought that the hydrosphere emerged as the result of processes taking place in the lithosphere, which has released a substantial quantity of water vapour and juvenile waters during the Earth's geological history.

Available data on the Earth's overall water balance are insufficient to permit studies of fluctuations in that balance in the geological past. Data concerning the redistribution of water between different components of the water balance during the appearance and disappearance of major glaciations are more reliable.

Paleogeographical data indicate that the level of the World Ocean declined by more than 100 meters during the age of the greatest Quaternary glaciations. Should currently existing ice sheets melt completely, the ocean level would increase by 66 meters. Such changes in the ocean level result in several per cent reductions or else increases in its area.

The ocean includes approximately ninety-five per cent of the total volume of free water. Ocean waters contain a substantial quantity of salt (35 grams per liter on the average). Their salinity increases somewhat in the subtropical high-pressure belt, where evaporation from the surface of the oceans exceeds precipitation, and declines near the Equator and also at middle and high latitudes, where precipitation exceeds evaporation. Sodium chloride and magnesium chloride constitute a large part of the dissolved salts (78 per cent and 11 per cent respectively).

Of the gases that are dissolved in the oceanic waters oxygen and carbon dioxide are especially important for organisms. The quantity of oxygen in oceanic waters changes within wide boundaries, depending on temperature, the activity of living organisms, and certain other factors. Concentrations of carbon dioxide in the ocean also change, but the overall mass of CO_2 in oceans exceeds its quantity in the atmosphere by approximately sixty times.

Carbon dioxide in oceanic waters is assimilated by autotrophic organisms in the course of photosynthesis. Part of the carbon dioxide which enters into the organic matter cycle is expended on the construction of calcium skeletons and shells of various species; following the death of such organisms and the dissolution of their skeletons, it partly returns to the ocean, but some quantity is deposited on the bottom of the ocean in the form of carbonate sediments.

Aside from oxygen and carbon dioxide the life processes of autotrophic organisms require a variety of minerals, among which phosphates (salts of phosphoric acid) are especially important. The concentration of minerals in various parts of the ocean varies considerably. Phosphate concentrations are lowest at low latitudes, increase somewhat at middle latitudes, and are highest in polar regions.

Oceanic waters are continually in motion, which is largely due to differences in the temperature of the ocean's surface layers at various latitudes and to dynamic interaction of moving air with the surface of the ocean.

The surface layer of oceanic waters is highly mixed under the influence of wave motion, and the speed of currents is highest in this layer. At depths greater than several hundred meters the speed of motion of waters is usually not high and the mixing of waters is small. The temperature of oceanic waters at low latitudes declines with depth and reaches 1-3°C at depths below two kilometers. At high latitudes the vertical temperature gradient is small and at all depths the temperature just exceeds the freezing point of salt water.

Large areas of the ocean in the Arctic and Antarctic regions are covered with ice fields whose average thickness is several meters. The area covered by ice changes in the course of a year, attaining a maximum in the spring and declining to a minimum at the end of the summer.

A large part of surface waters on land is found in lakes, of which the largest is the Caspian Sea. Approximately one half of lake water is fresh water.

One of the major hydrological processes occurring on land is the formation of river run-off which compensates the movement of water vapour from water bodies to the continents.

Deposits of moisture on the upper layers of the soil are of major importance for living nature.

Without soil moisture autotrophic plant covers could not exist. Soil moisture also influences the type of vegetation. For example, abundant moisture and subsequently insignificant amount of air in soil produce the type of plant life that is characteristic of swampland; the slightly excessive or else optimal moisture conditions produce forests; moderately insufficient moisture produces savannas and steppes; and acute shortages of moisture produce deserts. In its turn the type of plant cover influences many other properties of natural zones.

Land waters exert a major influence on the upper layers of the lithosphere and are a major factor in physical and chemical erosion processes. Similarly continental glaciation processes exert a great influence on the Earth's surface, as characteristic forms of glaciation landscapes develop following the retreat of ice.

In conclusion let us note once more that physical and chemical processes in the hydrosphere are closely linked to similar processes in the atmosphere. The transformation of energy, the circulation of water, carbon dioxide and a number of other components of air and water take place within both the hydrosphere and atmosphere as a single system.

Soils. While in the atmosphere and hydrosphere organisms are found within layers of considerable thickness that are measured in kilometers, most organisms existing in the lithosphere are concentrated within a layer of soils whose depth generally does not exceed several meters.

Soils are formed from mixtures of mineral substances that have appeared as a result of the weathering of rocks, and organic substances which are the products of life activities and decomposition of organisms, especially of plants. These products partly accumulate at the soil surface (leaves, dry branches, etc.) and partly at certain depths (for example, dead roots). An important role in the process of decomposition of organic matter entering the soil is played by a variety of microorganisms, and also by actinomyces, fungi and various soil animals.

As a result of the operation of these factors in the soil, deposits of carbon, nitrogen, phosphorus, potassium, magnesium and other elements appear in forms that can be assimilated by higher plants. In this way soils exert a major influence on the productivity of plant covers.

Another important outcome of the process of the decomposition of organic matter in soil concerns the generation of carbon dioxide, which is added to the volume of carbon dioxide in the atmosphere and the hydrosphere, thus compensating for its utilization in photosynthesis.

Soils usually consist of several layers that often differ in colour. The upper layer (Horizon *A*) contains products of the decomposition of organisms and of a synthesis of organic and mineral substances. The next layer (Horizon *B*) consists primarily of mineral components. Flows of water moving from layer *A* to layer *B* carry dissolved mineral substances. A third layer (Horizon *C*) consists of substances produced by the disintegration of the basic rocks which change but little with time. Sometimes they are formed where they are located, at the site of rock disintegration, while at other times they are brought over by water flows, ice formations, or winds. In particular, particles carried by the wind produce deposits of loess which sometimes are very thick and highly fertile.

Various types of soils develop under the influence of the prevailing climate, the nature of the basic rock, plant cover, and soil organisms. The first two factors may be viewed as independent of the process of soil formation, while the next two are influenced substantially by that process. The influence of climatic conditions on types of soil existing in various regions is particularly great.

The major types of zonal soils at high and middle latitudes include tundra soils, podzols of coniferous forests, brown and grey soils of deciduous forests, chernozems, chestnut-brown soils of steppe regions, and brown and grey-brown soils of semi-deserts and deserts. In subtropical and tropical areas the principal types of soils are similar to those of middle latitudes, while in the humid regions of low latitudes one generally finds either red-yellow or else dark lateritic soils, and in the case of dryer regions, red-brown soils. All zonal types of soils correspond to specific types of plant cover, and the nature of the soils influences greatly the structure and physiology of plants. Local soil characteristics that derive from the properties of basic rock and from local moisture conditions also exert an important influence on plant life.

Living Organisms. At the present time approximately 500,000 species of plants are known, as well as 1,500,000 species of animals (including about 1,000,000 species of insects).

The distribution of living organisms within the biosphere is very uneven. They are largely concentrated near the Earth's surface. In the oceans they are concentrated in the upper layers of water which are reached by solar radiation, and on land in the soil and on its surface. The quantity of biomass per unit volume of the atmosphere is negligible above the Earth's plant cover, and it is also very small in deep areas of water bodies.

But there are also vast regions covered by ice or else by desert on the Earth's surface, where the quantity of living organic matter per unit area is either negligible or practically equal to zero. There one may find only accidental living organisms but not regular forms of life.

In our age nearly the entire primary productivity of living organisms is attributable to the photosynthesis of autotrophic plants for, by comparison, chemosynthesis provides an inconsiderable quantity of organic matter.

The energy transformations of various living organisms may be generally represented in the form of an energy pyramid whose base is the energy that autotrophic plants assimilate in the course of photosynthesis; that energy is contained in the organic matter of plants. A large part of that organic matter is expended on the life processes of the plants themselves while another part is assimilated by animals feeding on such plants.

The second layer of the energy pyramid presents the flow of energy from autotrophic plants to animals and heterotrophic plants who utilize it. Normally not more than 10 per cent of the energy that is assimilated at the lower level is transferred to living organisms consuming the energy of autotrophic plants. In the course of their own life activities these organisms also lose much energy for respiration and other functions.

The third level of the pyramid corresponds to the flow of energy to animals not from the primary energy of autotrophic plants but from the energy already assimilated by other animals. That level includes carnivorous animals.

Many living organisms are located simultaneously at the second and third levels of the energy pyramid. This includes man, who consumes part of the energy that is produced directly by plants and part of the energy that is produced by animals.

The energy pyramid reflects the close dependence of animals on plants, for the only source of energy that supports the life activities of all animal species derives from autotrophic plants, and this trophic linkage greatly influences the bodily structure of animals and their ecology.

The energy resources that are produced by the activities of autotrophic plants determine the energy basis available to animals in different natural regions, and hence the geographical distribution of animals and the size of their populations.

The energy pyramid reflects the transformation of the energy that is found in living organisms. That process is continued through the transformation of the energy contained in the biomass of dead organisms and in products of life activities of organisms. The transformation of that energy proceeds until it is fully expended following the mineralization of organic matter. Many organisms contribute to that process, including a variety of microbes.

In fact transfers of energy from some organisms to others occur in ways that are far more complex than is suggested by such a simplified scheme.

In describing the trophic interactions that exist in nature, one must take into account the conception of trophic or food chains. These encompass groups of organisms that transfer the energy of their biomass to others in a certain sequence. While there are usually more than three links in a food chain, they cannot be very numerous, since generally each successive link in the chain possesses approximately ten times less energy than the preceding one.

Accordingly, should the number of links in a food chain be ten, for example, the quantity of energy that is assimilated by the last link will be billions of times smaller than that in the organisms of the first link. This implies a corresponding reduction in the biomass of organisms consuming that energy, so that for chains of any significant length the biomass of the last link becomes negligible.

Since heterotrophic plants and animals often feed on several types of food, the food chains are parts of complex webs of relations linking different organisms with each other rather than isolated lines. Occasionally these relations may even be bilateral. It has already been noted that the distribution of living organisms within the biosphere is uneven. In part this reflects a still more uneven distribution of the primary productivity of organic matter in space and time.

It is not in all parts of the biosphere that life exists as an autonomous system, i.e., living organic matter may be produced not in all localities where organisms are found. It has already been noted that in the present age nearly all primary productivity derives from the activity of autotrophic plants. These plants create organic matter only on land surface and in the upper layers of water bodies, because photosynthesis does not take place in the atmosphere or in deep layers of the ocean and of the soil.

What is more, there are wide regions on the surface of continents and oceans with long seasons, in which the process of photosynthesis does not occur. Autotrophic plants cannot exist in many regions of continents because of insufficient heat or moisture. The photosynthetical activities of plants are also limited over a large part of the surface of the oceans by insufficiency of mineral substances. As a result these zones are practically barren.

Thus life's energy pyramid does not operate on the Earth's entire surface. In spite of billions of years of evolution autotrophic organisms have not been able to adapt to conditions that exist in a large part of our planet's external cover.

The decisive role of the environment on the activities of living organisms is illustrated by the phenomenon of geographical zonality discovered by Dokuchayev.

Depending on prevailing heat and moisture conditions natural plant cover produces a number of characteristic types. These include tundras, the coniferous forests of high and middle latitudes (taiga), mixed and deciduous forests of middle latitudes, subtropical and equatorial rain forests, the deciduous forests of the tropics, forest steppes, savannas, steppes, semi-deserts, and deserts. The fauna of the corresponding regions is closely linked to these types of plant cover, and its own zonal distribution is determined by climatic factors.

The following interesting question arises: To what extent can organisms alter their environment in ways that are desirable to them? Leaving aside the problem of the influence of organisms on the evolution of the biosphere, which will be considered later, let us note that the ability of plants and animals to regulate their environment is very limited.

The plant covers exert a certain influence on microclimate, changing air temperature and moisture content in lower layers and reducing wind velocity near the Earth's surface. At the same time, many types of plant cover influence soil moisture and the rate of evaporation as well as components of the energy balance of the Earth's surface.

Changes in microclimate in the plant cover are a direct consequence of the plants' physiological functions. Accordingly, it cannot be said that the influence of the plant cover on microclimate emerged in the course of evolution as a form of adaptation of plants to the environment. This may be seen from the fact that certain changes in microclimate produced by plants (such as increases in total evaporation) are sometimes unfavourable to the plant cover. But generally many characteristics of the microclimate within the layer of the Earth's plant cover contribute to the survival of plants notwithstanding fluctuations in external conditions. This is illustrated by the prolonged preservation of ancient forests in some regions in which altered climate prevents their renewal if they are felled.

It is interesting to note the influence of plant covers on the preservation of soils. In many cases these are rapidly disintegrated by erosion processes following the destruction of plants. Examples may be found in a number of regions of insufficient humidity. The plants of dry steppes or savannas that are destroyed through excessive grazing, fires, or other anthropogenic influences are renewed very slowly or else not renewed at all.

Finally one should note that the influence of plants on their environment is largely of local character.

The correspondence of types of plant cover to climatic conditions points to the determining influence of global climate on the development of plant life, and to the limited character of the plants' ability to regulate their environment. A similar conclusion may be drawn from the widely-known sensitivity of plants to major anomalies in weather and climate (frost, droughts, etc.).

While plant covers do influence the climate of large areas, that influence is relatively modest. Expenditures of solar energy on photosynthesis and on the generation of heat through the oxidation of organic matter generally do not exceed a fraction of a per cent of available solar radiation, and these two magnitudes often compensate each other.

Of all biological processes it is the plant cover's transpiration that influences the energy balance at the Earth's surface most. Expenditures of heat on transpiration may represent several tens per cent of the radiation balance at the Earth's surface. Yet the influence of transpiration on the heat balance and the thermal regime is not as large, since in the absence of transpiration an increase in evaporation from soil surfaces takes place.

The vegetation, and especially forests appear to exert some influence on the volume of precipitation. This is due to the intensification of the water exchange over forests through increased evaporation and to the increased roughness of the Earth's surface which intensifies vertical air currents. As a result the total precipitation is somewhat increased, but not significantly. This process is considered more closely in Chapter IV.

Plants may influence the rate of absorption of solar radiation. On land, thick plant covers usually possess less albedo than do less developed plant growths. Similarly the albedo of water bodies with a strong concentration of water plants is also reduced. While all these factors influence the climate to some extent they do not produce substantial changes in global climate.

The influence of animals on the environment is even less pronounced. It is most perceptible in the case of animals living in the soil that often alter its structure. But such changes are of a local character.

The Evolution of Organisms. The Earth's flora and fauna have developed over a major part of its geological history. Some data relating to distinct ages within that historical process are presented in the Geological Time Scale.

The scale is based on successive changes in the nature of lithogenesis, and in the flora and fauna which have been established through studies of the Earth's crust.

Data relating to fossil organisms decline rapidly as we proceed deeper into the past. Even though the Archeozoic and the Proterozoic eras were several times longer than the Phanerozoic eon (which includes the Paleozoic, Mesozoic and Cenozoic eras) the volume of data on Pre-cambrian organisms is small compared with organisms of the Phanerozoic eon.

It is believed that autotrophic plants have appeared 3-3.5 billion years ago. The extreme paucity of remains of plants and animals from the Pre-cambrian periods is probably due to the weak development of hard tissues in these organisms, and also, perhaps, to the complete vanishing of such ancient remains of organisms through the metamorphism of rocks.

Most of the data pertaining to Archeozoic organisms are of indirect nature. For example, carbonaceous deposits of the Archeozoic era could have appeared as a result of the activities of bacteria and seaweeds absorbing mineral substances of oceanic waters. In the case of the Proterozoic era, various remains of water organisms have been found, includ-

THE GEOLOGICAL TIME SCALE

Era	Period	Epoch	Duration of period (millions of years)	Absolute age of beginning of period (millions of years)
Cenozoic	Quaternary	Holocene Pleistocene	2	2
	Tertiary	Pliocene Miocene Oligocene Eocene Paleocene	64	66
Mesozoic	Cretaceous		66	132
	Jurassic		63	185
	Triassic		50	235
Paleozoic	Permian		45	280
	Carboniferous		65	345
	Devonian		55	400
	Silurian		35	435
	Ordovician		55	490
	Cambrian		80	570
Proterozoic				
Archeozoic				

ing blue-green seaweeds, and also various animals, including (in the case of the upper Proterozoic era) such multicellular organisms as coelenterata.

During the Cambrian period there existed many water animals, including hundreds of types of trilobites, brachiopoda, archeocyte, crustacea, and various molluscs and medusae. The plant life of the Cambrian period included various seaweeds.

In the Ordovician and Silurian periods the number of types of invertebrate animal increased substantially, and together with trilobites many types of arthropoda became widespread, including ancient crawfishes and scorpions. Some of the arthropoda became the first inhabitants of dry land. It is at that time that the first vertebrates appeared. These were primitive, jawless fish. In the Ordovician and Silurian periods the land's plant cover emerged, which included various psilophytes.

The Devonian period was characterized by major changes in the composition of the flora and fauna. It was at that time that crossopterygii fish appeared, followed by amphibians which derived from crossopterygii fish. Insects first came into existence at the end of the Devonian period. Together with new groups of animals new plants also appeared, ancient psilophytes were replaced by equiseta, ferns, and gymnospermous plants. The flora reached its height during the Carboniferous period, when large areas came to be covered by forests whose remains subsequently turned into deposits of coal.

During the Carboniferous period many forms of land insects emerged, some of which were gigantic. The first reptiles appeared towards the end of the Carboniferous period.

The Permian period was characterized by a wide dissemination of gymnospermous plants (cycadaceae and ginkgoaceae) that became dominant towards the end of that period. There were many amphibia and reptiles who developed rapidly.

The organic world changed substantially at the threshold of the Mesozoic era. Many types of sea invertebrata vanished, including trilobites, and the numbers of others declined.

The Triassic period marked the height of development of reptiles. This lasted throughout the entire Mesozoic era.

During the Triassic period turtles and dinosaurs appeared, as well as sea reptiles—ichthyosaurs and plesiosaurs. The first primitive mammals came into existence at the end of of the Triassic period.

During the Jurassic period giant amphibians vanished but many other, even larger reptiles appeared, including various dinosaurs. Flying reptiles, pterodactyls, became numerous. At the same time the first birds appeared (archeopteryx) which were in many ways close to reptiles. Together with ancient fish—sharks and skates— the first bony fish made their appearance. The fauna of that time, as of the subsequent Cretaceous period, did not include many mammals.

During the Cretaceous period snakes appeared, as well as new species of dinosaurs and large flying reptiles (pterodonts).

Throughout most of the Mesozoic era coniferous plants and ferns prevailed on the continents. Towards the middle of the Cretaceous period the first angiospermae emerged and spread rapidly on all continents, producing forests of oak, poplar, beech, palm, and other trees.

The beginning of the Cenozoic era marked an important boundary in the history of the organic world. Towards the end of the Cretaceous period the overwhelming majority of groups of reptiles vanished, including land dinosaurs, sea reptiles—ichthyosaurs and plesiosaurs, and flying reptiles. Of the various reptiles of the Mesozoic era only crocodiles, turtles, lizards and snakes remained.

At the beginning of the Tertiary period the evolution of the placental mammals that appeared during the Cretaceous period sharply intensified. Over a short period of time mammals occupied ecological niches that were vacated by the extinct reptiles, and many of them became large and even gigantic. Some groups settled in the ocean while others acquired a capacity to fly. Together with mammals, birds came to play a large role in the fauna of the Cenozoic era. They also partly filled the ecological niches that had been formerly occupied by reptiles.

Throughout the Cenozoic era plant forms changed relatively little, although zones with different types of plant cover migrated over large distances because of changes in climate, particularly during the Quaternary period.

The end of the Cenozoic era was marked by the appearance of man. Even though Primates closely related to man had already existed in the Paleocene epoch, for a long time their evolution was relatively slow. In the Lower Oligocene the first anthropoid apes appeared, while in the Pliocene, the first Australopithecus close to man arose, and then the first Hominidae. Modern man appeared during the last ice age, 35-40 thousand years ago, and his activities have exerted an enormous influence on the evolution of the biosphere.

The Self-Regulation of the Biosphere. The biosphere appeared together with the emergence of life on Earth, that is, about three billion years ago. The prolonged existence of the biosphere is often regarded as something so natural that the need to explain this fact is not recognized. Yet all components of the biosphere, and especially living organisms possessing limited measures of stability in the event of environmental changes, could easily have vanished in certain combinations of external factors.

The preservation of living organisms over billions of years was possible only because of the relatively small changes in their environment. It is known that the active existence of the overwhelming majority of organisms is possible only within relatively narrow boundaries of climatic conditions (temperatures ranging from 0 to 50° C, the presence of accessible forms of water, oxygen, etc.). Autotrophic plants can develop only given specific concentrations of carbon dioxide in the air or in the water.

The relative stability of climatic conditions needed for the prolonged existence of life results from the operation of several linkages between elements of meteorological processes. This problem will be considered in more detail in Chapter III, and now let us note the most important type of feedback relations that maintain the stability of the Earth's climate, namely, the dependence of the long-wave radiation emitted into outer space on the temperature of air and of the Earth's surface. As the inflow of heat increases so does that temperature, and this increases the outgoing radiation. This produces a partial compensation of increased inflows of heat, thus attenuating corresponding increases in temperature.

Such a feedback relation increases the stability of the Earth's thermal regime despite oscillations of the inflow of heat from outer space and represents a negative feedback relation.

Negative feedback relations also exist that increase the stability of individual elements of the hydrosphere. The stability of closed water bodies, for example, depends to a large extent on the sloping shape of their shores. As the inflow of water increases so does the water level and the area covered by the body of water, and this increases the volume of surface evaporation, which in turn reduces further increases in the water level.

It has been shown that in the case of closed bodies of water with vertical shores, where there is no such a negative feedback relation, large oscillations in the water level are possible, up to the complete disappearance of the water body from time to time, even in the case of relatively stable climatic conditions.

The stability of the World Ocean is explained by more simple factors: the volumes of all other components of the biosphere waters are relatively small, and this makes substantial fluctuations in the volume of oceanic waters impossible.

Turning to the problem of the autoregulation of plant and animal populations, we must note that the number of various organisms may fluctuate widely even when external conditions remain constant, because of the struggle for existence between species and other biological factors. Changes in the numbers of individual species of plants and animals are checked by many negative feedback relations that have appeared in the course of the evolution of organisms. This will be considered more closely in Chapter VII; here we shall only mention a relationship that is frequently the most important one: as the numbers of a particular species of organisms begin to exceed a certain upper threshold, the intraspecies struggle for existence increases hindering the further growth of numbers. The opposite process arises when numbers decline.

Much less is known about the autoregulation of the biosphere as a whole than about the autoregulation of its individual components.

It is clear that feedback relations exist among individual components of the biosphere, some of which are negative and provide for the stability of the entire system. Nevertheless, the forms of these feedback relations are not well known.

One should expect that the dependence of the composition of the air in the atmosphere on the activities of living organisms is substantial. Since that activity depends greatly on the state of the atmosphere this is a feedback relation which may be described in the following schematic form.

For moderate concentrations of carbon dioxide the productivity of the autotrophic plant cover is approximately proportional to that concentration. The major part of the output of autotrophic plants is oxidized, and as a result the carbon dioxide assimilated by plants is returned into the atmosphere and the hydrosphere while the oxygen released as a result of photosynthesis is absorbed in the course of the decomposition of organic matter. Yet the biological cycle of carbon dioxide is not fully closed, and a part of the organic matter produced by plants in a variety of ways is sedimented. Assuming that that part is proportional to the total productivity of autotrophic plants, we may conclude that the activity of plants produces a continuous expenditure of free carbon dioxide and also an inflow of free oxygen that are proportional to the concentration of carbon dioxide.

The inflow of biological oxygen is added to its expenditure on the oxidation of mineral substances, which increases as the concentration of oxygen increases. This produces a negative feedback relationship that regulates the quantity of free oxygen. That quantity also depends on the concentration of carbon dioxide, which is determined by the inflow of that gas from the Earth's crust, its expenditure within the biological cycle, and the outflow associated with the formation of carbonates and the weathering of rocks. Since both forms of expenditure of carbon dioxide increase as its concentration increases, we find that the quantity of free carbon dioxide is also regulated by a negative feedback relation.

The carbon dioxide and oxygen balance will be considered in more detail in later chapters; here let us only note that the mechanism we have described is probably important for the longterm preservation of free oxygen and carbon dioxide in quantities required for the existence of autotrophic plants.

In conclusion let us note that the biosphere is a complex system that contains an hierarchy of systems of varying degrees of complexity. The highest position within that hierarchy is occupied by systems corresponding to individual components of the biosphere (the atmosphere, hydrosphere, soil, and living organisms), each of which may be divided into more or less complex systems. At the same time a diversity of complex ecological subsystems of various sizes may be identified within the biosphere.

Systems included in the biosphere possess a definite measure of stability that is provided by a complex of internal feedback relations within these systems as well as by interrelations among different systems.

Such a structure of the biosphere makes it possible to study it using methods of systems analysis.

It is clear that the study of problems of global ecology call for the study of the large-scale systems that relate to the biosphere as a whole, or else to substantial portions of the biosphere. In addition an understanding of the patterns of global ecological processes frequently requires studies of local systems in specific regions.

II

The Transformations of Solar Radiation

1. Components of the Biosphere's Energy Balance

The Transformations of Solar Energy. Solar energy is the only source of energy for nearly all natural processes occurring in the biosphere.

It was Alexander Ivanovich Voeikov, a prominent Russian climatologist, who first understood the importance of transformations of solar energy in studying the biosphere. Writing at the end of the nineteenth century he observed: "I believe that one of the major tasks of the physical sciences today is the initiation of a book of accounts of incomes and expenditures of the solar heat that is received by the Earth and its envelope of air and water" (1884).

The problem formulated by Voeikov has been tackled by many researchers in various countries, and important contributions have been made by W. Schmidt, A. Angström, H. Sverdrup, S. I. Savinov, N. N. Kalitin, and F. Albrecht.

In the USSR systematic studies of the Earth's energy (heat) balance were carried out since 1945 in Leningrad Voeikov Main Geophysical Observatory, where many maps of the energy balance components have been compiled.

In estimating the specific components of the energy balance these studies have relied on data from actinometric observations and on calculations based on semi-empirical formulae that relate these components to meteorological data collected by the established network of stations. More recently studies of the energy balance have also made use of data received from Earth meteorological satellites.

The general pattern of the transformations of solar energy that occur in the biosphere may be described as follows.

At the mean distance between the Earth and the Sun the flux of solar radiation is approximately 1,000 kcal/cm^2·year. Because of the Earth's spherical shape one-fourth of the overall volume of that flux, i.e., 250 kcal/cm^2·year enter a unit of surface at the atmosphere's upper boundary, and approximately 170 kcal/cm^2·year are absorbed by the Earth itself.

As it is heated through the absorption of solar energy the Earth's surface becomes a source of long-wave radiation that heats the atmosphere. The water vapour and various gases that are contained in the atmosphere and that absorb long-wave radiation decrease the flux of radiation from the Earth's surface. As a result a substantial part of the radiation of the Earth's surface is compensated by the atmospheric counter-radiation.

The difference between radiation from the Earth's surface itself and the atmospheric counter-radiation is called effective radiation. The magnitude of effective radiation is usually several times smaller than the flux of long-wave radiation from the Earth's surface that would be observed were the atmosphere fully transparent to long-wave radiation.

The algebraic sum of energy radiation fluxes reaching the Earth's surface and leaving it is called the Earth's surface radiation balance. It is evident that the magnitude of the radiation balance is equal to the difference between the quantity of direct and diffused short-wave radiation absorbed by the Earth's surface and the quantity of effective long-wave radiation.

For the Earth's surface as a whole average effective radiation is substantially lower than absorbed short-wave radiation. This follows from the so-called greenhouse effect, i.e., results from the relatively greater transparency of the atmosphere to short-wave radiation than to long-wave radiation. The Earth's surface average radiation balance is therefore a positive value.

The energy of the Earth's surface radiation balance is expended on heating the atmosphere through turbulent heat conductivity, on evaporation, and on heat exchange with deeper layers of the hydrosphere and the lithosphere, etc. The quantitative characteristics of all forms of transformation of solar energy on the Earth's surface are represented in the equation of the energy (heat) balance. This includes the algebraic sum of flows of energy reaching the Earth's surface and those leaving it. In accordance with the law of conservation of energy that sum is equal to zero.

The radiation and the heat balances at the Earth's surface are linked to the corresponding balances within the atmosphere.

The Earth as a planet receives heat from outer space and returns it to outer space exclusively in the form of radiation. Since the Earth's average temperature changes but little over time, it is evident that its radiation balance (the difference between absorbed radiation and radiation emitted into outer space) is equal to zero.

This is why the atmosphere's average radiation balance, equal to the difference between the Earth's radiation balance and that of the Earth's surface, is a negative magnitude equal to the absolute value of the radiation balance at the Earth's surface. The negative radiation balance of the atmosphere is compensated by inflows of energy from the condensation of water vapour associated with the formation of clouds and precipitation, and by the flow of heat from the Earth's surface that is associated with the turbulent heat conductivity of the air lower layer.

In addition to processes of vertical redistribution of solar energy within the atmosphere and the oceans powerful horizontal redistribution processes exist as well. One of them relates to the transfer of heat energy from low latitudes to higher ones as a result of the uneven heating of the Earth's spherical surface by radiation. That transfer is effected in the form of macroturbulent heat exchange and transfers of heat through well-ordered movements, as well as redistribution of condensation heat (in the atmosphere).

In their turn these processes of transformation of solar energy which are produced by radiation factors alter substantially the radiation regime itself, which largely depends on circulation processes in the atmosphere and in the oceans and on changes in the state of water. In particular it is strongly influenced by cloudiness and by snow and ice covers on the Earth's surface.

In addition to processes of transformation of solar energy of "the first order", which change the radiation and heat processes within the biosphere substantially, there also occur a number of transformations of solar energy that result from expenditures of relatively small quantities of heat and therefore do not exert a perceptible direct influence on the radiation and heat regimes. Yet some of these processes are of very great importance to other components of the biosphere. This is true, for example, of the process of photosynthesis, which transforms radiation energy into a relatively stable form of chemical energy and creates organic matter.

The principal data that are employed in studying transformations of solar energy in the biosphere relate to the energy balance. The data on the energy balance of the Earth's surface are especially important since the latter is the biosphere's major source of energy and the greater part of the biomass of living organisms is concentrated there.

Radiation Balance. The Earth's surface radiation balance R is equal to the difference between the absorbed solar radiation and the long-wave effective radiation:

$$R = Q(1-\alpha) - I, \qquad (2.1)$$

where Q is the total short-wave radiation (the sum of direct and diffused radiation); α is the albedo (the reflectivity of the Earth's surface in relation to total radiation) expressed in terms of a fractions of a unity; and I is the effective radiation, i.e., the difference between the Earth's surface radiation and the atmospheric counter-radiation that is absorbed at the Earth's surface.

The radiation balance of the Earth-atmosphere system may be determined in a similar way, where R_s represents the radiation balance of a vertical column passing through the entire atmosphere until it reaches the Earth's surface.

In such a case we obtain

$$R_s = Q_s(1-\alpha_s) - I_s, \qquad (2.2)$$

where Q_s is solar radiation reaching the atmosphere's upper boundary; α_s is the system's albedo; and I_s is a long-wave radiation from the atmosphere's upper boundary into outer space (outgoing radiation).

For the atmosphere the radiation balance R_a is the difference between the magnitudes of R_s and R. Making use of formulae (2.1) and (2.2) we obtain

$$R_a = Q_s(1-\alpha_s) - Q(1-\alpha) - (I_s - I). \qquad (2.3)$$

At the mean distance between the Earth and the Sun the flow of solar energy has been estimated at 1.95 cal/cm²/min.

The maximum radiation within the solar spectrum occurs within its visible range (for wave lengths ranging from 0.40 to 0.75 micrometer). In addition an important quantity of energy is transmitted in the form of infrared radiation, with wave lengths of more than 0.75 micrometer. Ultraviolet radiation (with wave lengths of less than 0.40 micrometer) is relatively small, but its influence on various biological processes is considerable.

An appreciable part of the radiation from the Sun is absorbed and diffused within the atmosphere and is also reflected back into outer space. Clouds exert a particularly strong influence on the scattering of solar energy, but even in the absence of clouds solar radiation in the atmosphere changes substantially.

The total solar radiation reaching the Earth's surface consists of direct radiation and diffused radiation.

The reflection of solar radiation from the Earth's surface is characterized by the magnitude of the albedo, which is equal to the ratio of the reflected radiation to the incoming radiation. In theoretical terms values of the albedo may range from one, for absolutely white and fully reflecting surfaces, to zero, for absolutely black surfaces that fully absorb solar rays.

Data supplied by observations indicate that the albedo values of natural underlying surfaces vary widely, and that their changes cover nearly the entire range of possible values of reflectivity of different surfaces.

Experimental studies of reflection of solar radiation have established albedo values for all more or less widely-occurring natural underlying surfaces. These studies have shown that there is a considerable difference between the conditions of solar radiation absorption on land and waters.

The largest albedo values, of 0.90-0.95, were recorded for pure and dry snow. As the snow cover is seldom pure, the average albedo values for snow are usually lower and range from 0.70 to 0.80. The albedo of wet or dirty snow may be as low as 0.40-0.50.

In the absence of snow the greatest albedo values at the Earth's surface were found in certain desert regions in which the surface of the soil is covered by a layer of crystalline salts (dried salt lakes). In such cases the value of the albedo may reach 0.50. It is somewhat less in desert areas that do not possess any plant life and consist of light sandy soils.

It should be noted, however, that the albedo of desert regions may vary within a wide range, depending on the colour of the soil.

The albedo of damp soil is usually less than the albedo of the analogous dry soil. For damp chernozems the albedo declines to very low values and may reach 0.05.

The albedo of natural surfaces with thick vegetation varies within relatively narrow range, generally from 0.10 to 0.20-0.25. Observations show that the albedo of forests (especially coniferous forests) is generally less than that of low plant forms characteristic of meadows. This appears to be due to more favourable absorption conditions in the case of the tall vegetation cover, where there is a higher probability of absorption of solar energy fluxes following their reflection by plants.

In the water bodies conditions of absorption of solar radiation differ from those on land. Water that is more or less clean is relatively transparent to short-wave radiation. As a result solar rays that penetrate the upper layers of such a body of water are repeatedly dispersed and then largely absorbed.

It is not difficult to imagine that the process of absorption of solar energy in these conditions depends substantially on the height of the Sun. When the Sun is relatively high a substantial part of the incoming radiation penetrates the upper layers of the body of water and is largely absorbed. When the Sun is low its rays reach the water surface at low angles, are reflected, and do not reach great depths. This causes the albedo to increase sharply.

It has been established that when the Sun's position is high the albedo of its direct radiation does not exceed several hundredths, while as it approaches the horizon that value becomes several tenths.

In the case of diffused radiation the albedo of a water surface appears to be much less variable and is approximately 0.10.

The albedo of the Earth-atmosphere system is more complex in nature than the albedo of the Earth's surface.

The solar radiation reaching the atmosphere is partly reflected as a result of the back scattering of the atmosphere. When there are clouds, a substantial part of such radiation is reflected from their upper surfaces. When there are few clouds, or else no clouds at all, the albedo of the

Earth-atmosphere system depends substantially on the albedo of the Earth's surface. The albedo of clouds depends on their thickness, and its average value is 0.40-0.50. In the presence of clouds the albedo of the Earth-atmosphere system is usually larger than that of the Earth's surface, except in those cases when the Earth's surface is covered with relatively clean snow.

In addition to the reflection of short-wave radiation, long-wave radiation is an equally important form of expenditure of radiation energy.

The basic patterns of the heat transfer through long-wave radiation may be described as follows.

In accordance with the Stefan-Boltzmann Law, the radiation of an absolutely black body is equal to σT^4, where T represents temperature measured in terms of Kelvin's scale, σ is Stefan's constant (equal to $8.14 \cdot 10^{-11}$ cal/cm$^2 \cdot$min$\cdot$deg.^{4}). Available experimental data indicate that the radiation of the Earth's natural surfaces is generally quite close to the radiation of a black body at the corresponding temperatures, and that the ratios of the observed values of radiation to the radiation of a black body mostly fall within a range of 0.90-1,00.

The intensity of long-wave radiation from the Earth's surface at various wave-lengths depends on the temperature of the radiating surface. The major part of radiation energy occurs at wave-lengths that range from five to several tens of micrometers. A significant part of the flow of long-wave radiation from the Earth's surface is compensated by counter-radiation from the atmosphere.

In the absence of clouds the atmospheric long-wave radiation is largely determined by the presence of water vapour and carbon dioxide; the influence of atmospheric ozone is usually less important.

Water vapour and carbon dioxide absorb the long-wave radiation primarily in specific spectral regions. As a result the atmospheric long-wave radiation is also quite heterogeneous from a spectral point of view.

The most pronounced absorption of radiation by water vapour occurs in the range of wave-lengths from 5 to 7.5 micrometers while between 9 and 12 micrometers the absorption of radiation is relatively low. Carbon dioxide posseses several absorption bands among which the one ranging from 13 to 17 micrometers is especially important.

It should be noted that the carbon dioxide component of the atmosphere changes relatively little while the quantity of water vapour may be very different depending on meteorological conditions. Changes in the air humidity may substantially influence the magnitude of atmospheric radiation.

The atmospheric counter-radiation generally increases considerably in the presence of clouds. Clouds in the lower and middle layers are usually rather dense and at corresponding temperatures radiate as absolutely black bodies. Under such conditions effective radiation is largely determined by the difference in temperatures at the Earth's surface and at the lower surface of clouds. When that difference is not large effective radiation, too, approaches zero and may even become a negative magnitude.

High clouds usually radiate less than a black body, because of their low density. Since the temperature of such clouds is relatively low it may be concluded that their influence on effective radiation is small compared with clouds in lower and middle layers.

Thus the effective radiation of the Earth's surface depends largely on temperature at the Earth's surface, the air humidity, and clouds.

Outgoing emission of the Earth-atmosphere system encompasses that part of the radiation of the Earth's surface which passes through the atmosphere plus the radiation of the atmosphere itself. In the absence of clouds an important role within the system is played by the radiation of the Earth's surface within a wave-length range of 9 to 12 micrometers. In completely cloudy conditions the radiation from the clouds' upper surface acquires a major importance and depends on temperature at that surface. Since that temperature is usually much lower than the temperature of the Earth's surface clouds serve to reduce substantially the volume of heat that is returned by long-wave radiation into outer space.

The Energy Balance. Equations representing the energy balance constitute specific formulations of a law of physics, namely the law of conservation of energy. The equations may be compiled for various volumes and surfaces of the atmosphere, hydrosphere, and lithosphere. In studying the biosphere the equations are most often employed of the balance for specific sectors of the Earth's surface and of that for the Earth-atmosphere system, i.e., for a vertical column passing through the entire biosphere.

Equations of the energy balance of the Earth's surface include flows of energy between elements of the surface and the surrounding space. These flows include flows of radiation heat whose sum is equal to the radiation balance.

The positive or negative magnitude of the radiation balance is compensated by several flows of heat. Since the temperature of the Earth's surface usually differs from that of the air, a flow of heat develops between the underlying surface and the atmosphere which is determined by turbulent heat conductivity.

A similar flow of heat takes place between the Earth's surface and deeper layers of the lithosphere and hydrosphere whenever the temperature at the surface differs from that of deeper layers. The flow of heat within the soil is then determined by the mechanism of molecular heat conductivity, while the heat conductivity of water bodies is usually turbulent in various degrees.

The expenditure of heat on evaporation usually exerts a substantial influence on the heat balance of the Earth's surface. The magnitude of evaporation depends on the moistening of the Earth's surface, its temperature, the air, humidity and also on the intensity of turbulent exchange in the layer of air adjoining the Earth's surface which determines the rate of transfer of water vapour from the Earth's surface into the atmosphere. This last factor influences the patterns of expenditure of heat on evaporation which often prove to be similar to those of turbulent heat exchanges between the Earth's surface and the atmosphere.

Expenditures of heat on evaporation change their sign when the direction of the flow of water vapour changes. When water vapour moves from the atmosphere to the Earth's surface the expenditure of energy on evaporation is replaced by an inflow of heat resulting from the condensation of water vapour.

In composing an equation representing the energy balance let us denote the radiation flow of heat by R, the turbulent flow of heat from the underlying surface to the atmosphere by P, the flow of heat from the underlying surface to lower layers by A, and expenditures of heat on evaporation (or else the release of heat in condensation) by LE, where L is the latent heat of vaporisation and E is the rate of evaporation. Since all other variables in the heat balance are usually much smaller than the magnitudes of the flows of heat that have been mentioned, the energy balance may be roughly written as follows:

$$R = LE + P + A. \tag{2.4}$$

The magnitude of the radiation flux (radiation balance) R is considered to be positive when it describes the inflow of heat to the underlying surface; all other variables will be positive when they describe an expenditure of heat. (The flows of heat corresponding to the heat balance equation are represented in Figure 1.)

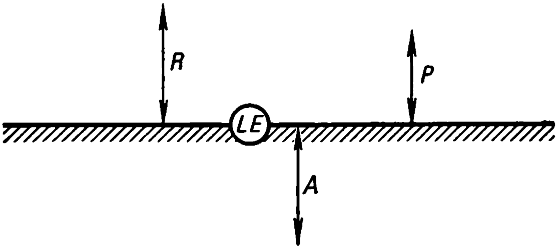


Fig. 1. The Earth's Surface Energy Balance

Among the elements of the energy balance that have not been included in equation (2.4) the expenditure of heat on the melting of ice or snow on the Earth's surface is the greatest (or alternatively, the inflow of heat from the freezing of water). Although for long periods of averaging (for instance, for a year) the latter magnitude is generally considerably less than the basic components of the balance, for some cases, however (for instance, for the snow-melting period at high and middle latitudes) the magnitude discussed should be included in equation (2.4) as an additional term.

Other elements of the energy balance include flows of heat from the dissipation of energy associated with the friction of air currents, waves produced by wind, tides; heat flows (positive or negative) that are transferred by precipitations whenever their temperature is not equal to that of the underlying surface; and also expenditures of energy on photosynthesis and inflows from the oxidation of biomass. For any considered periods their average magnitudes are substantially smaller than those of the basic components of the balance. Even though there may be exceptions to this rule (such as, for example, in the case of forest fires when a large quantity of heat that has been accumulated through photosynthesis over many years is suddently released) these are relatively rare, particularly for large sectors of the Earth's surface.

The magnitude of the flow of heat A from the Earth's surface to lower layers may be determined through other components of the energy balance of the upper layers of the lithosphere or hydrosphere (Fig. 2). If we should form an equation for the energy balance of a vertical column whose upper base is located at the Earth's surface and the lower base lies at depths where the flow of heat is negligible, we will obtain the equation

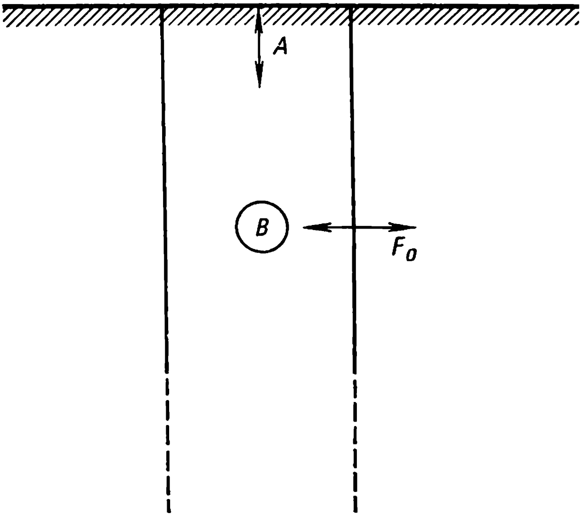


Fig. 2. The Energy Balance of the Upper Layer of the Lithosphere or Hydrosphere

$$A = F_0 + B, \tag{2.5}$$

where F_0 is the inflow of heat produced by horizontal heat exchange between the column and the surrounding space of the lithosphere or hydrosphere, and B represents changes in the heat content inside the column over the given period of time.

At the bottom of the column the vertical flow of heat may be assumed to be zero since the flow of heat from the depths of the Earth's crust is usually negligibly small by comparison with the basic components of the energy balance.

The magnitude of F_0 is equal to the difference between the flows of heat that enter through the vertical walls of the column under consideration and that leave it in a similar way.

The term F_0 becomes large in the case of large water bodies in which currents exist with a large horizontal heat conductivity that is determined by the operation of macroturbulence.

For conditions existing in the lithosphere the value of F_0 is usually negligible because of the low heat conductivity of soil. Accordingly, for land $A=B$. Since over a period of a year the upper layers of soil are neither cooled nor heated on the average, than for an average yearly period over many years on land $A=B=0$.

In the case of closed water bodies taken as a whole, whose area and depth are relatively large, the values of A in their energy balance are also close to those of B, since heat exchanges between such bodies of water and the ground are usually negligible compared with the values of the basic components of the heat balance.

Yet in specific sectors of the ocean and of other bodies of water (seas and lakes) the magnitudes A and B may differ substantially. In such cases the average yearly value of heat exchange of an active surface with lower layers is equal not to zero but rather to the quantity of heat that is either received or lost by a vertical column passing through the hydrosphere as a result of the influence of currents and of macroturbulence (i.e., $A=F_0$).

Thus for an average yearly period the heat balance equation for land will be

$$R=LE+P, \tag{2.6}$$

while for the ocean it will be

$$R=LE+P+F_0. \tag{2.7}$$

In certain situations both these equations may be further simplified. Thus for desert conditions, in which evaporation is close to zero, equation (2.6) acquires the form $R=P$. For the World Ocean as a whole, in which the redistribution of heat by currents is compensated and is equal to zero, equation (2.7). becomes $R = LE + P$.

In deriving the equation of the heat balance of the Earth-atmosphere system one should consider the inflow and expenditure of heat energy in a vertical column passing through the entire atmosphere and the upper layers of the hydrosphere or lithosphere down to levels at which noticeable seasonal (or else daily) fluctuations of temperature cease.

Heat exchanges between such a column and outer space are described by its radiation balance R_s, which is equal to the difference between the solar radiation that is absorbed by the entire volume of the column and the total long-wave radiation from that volume (Fig. 3). We will consider the magnitude R_s to be positive when it describes inflows of heat into the Earth-atmosphere system.

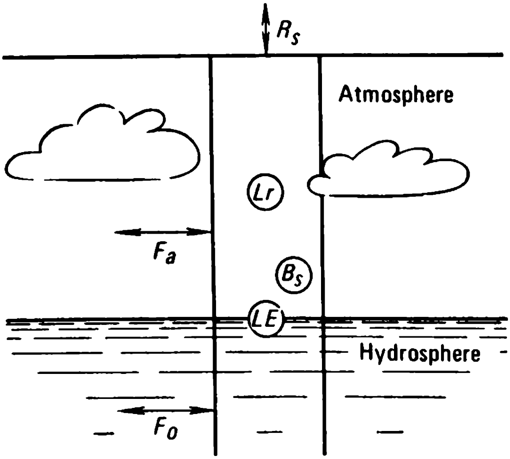


Fig. 3. The Energy Balance of the Earth-Atmosphere System

If one extends the column into the lithosphere or hydrosphere down to layers whose thermal state does not depend on changes in meteorological factors, it may be assumed that the flow of heat through its lower base is practically equal to zero.

Flows of heat through the sides of the column depend on horizontal transfers of heat in the atmosphere and hydrosphere. Differences in inflow and expenditure of heat resulting from transfers in the atmosphere are described on the diagram by the arrow marked F_a, and in the hydrosphere by the arrow marked F_o.

It should be noted that the value of F_a is physically similar to that of F_o. It describes the inflow or expenditure of heat in a column of air resulting from the atmospheric advection and macroturbulence.

Aside from heat exchanges through the surface of such a column its heat balance is influenced by sources of heat (both positive and negative) that are located within the column itself. These include the inflow and expenditure of heat resulting from changes in the state of water, especially as a result of evaporation and condensation.

The inflow of heat from condensation in the atmosphere is equal to the difference between the inflows and expenditures of heat associated with the condensation and evaporation of drops of water in clouds and fogs. Over surfaces that are sufficiently homogeneous during long periods the average difference in the magnitudes of condensation and evaporation in the atmosphere is equal to the sum of precipitations r, the inflow of heat being equal to Lr. The corresponding component in the energy balance represents the difference between heat inflow from condensation and its expenditure in the evaporation of drops. It may, however, differ from the magnitude Lr in the conditions of the rugged surface and also in individual short periods of time.

Expenditures of heat on evaporation from the Earth's surface (the difference between heat expenditures on evaporation from the surface of water bodies, soils and vegetation and inflows of heat from condensation on these objects) are equal to LE. The overall influence of condensation and evaporation on the column's energy balance may be approximately characterized in terms of the magnitude of $L\,(r-E)$.

Among other components of the column's energy balance one should consider B_s, i.e., the change in the heat content within the column over the period of time to which the summation refers. Remaining components of the balance (the inflow of heat from the dissipation of mechanical energy, the difference between heat expenditure on the melting of ice and the inflow of heat from its formation, the difference between heat expenditure on photosynthesis and the inflow of heat from the oxidation of organic matter, etc.) do not usually play a perceptible role in the heat balance of the Earth-atmosphere system and may be neglected.

In writing the equation for the energy balance of the Earth-atmosphere system as

$$R_s = F_s + L(E - r) + B_s, \tag{2.8}$$

where $F_s = F_a + F_o$, we will assume that all terms on the right-hand side are positive when they describe an expenditure of heat. For an average yearly period the magnitude of B_s is apparently close to zero, and equation (2.8) becomes

$$R_s = F_s + L(E - r). \tag{2.9}$$

For land conditions that equation is

$$R_s = F_a + L(E - r). \tag{2.10}$$

Since for the entire globe over a one-year period $E = r$, and the horizontal inflow of heat into the atmosphere and hydrosphere is apparently zero, the equation for the heat balance of the biosphere as a whole acquires the following simple form:

$$R_s = 0. \tag{2.11}$$

The equation representing the heat balance of the atmosphere may be obtained either by summing the corresponding flows of heat, or else as a difference between members in the heat balance equation for the Earth-atmosphere system and in that for the Earth's surface.

Assuming that the atmospheric radiation balance R_a is equal to $R_s - R$ and that changes in the heat content of the atmosphere B_a are equal to $B_s - B$, we find that

$$R_a = F_a - Lr - P + B_a, \tag{2.12}$$

and that for an average yearly period

$$R_a = F_a - Lr - P. \tag{2.13}$$

2. *The Geographical Distribution of Energy Balance Components*

World Maps. One may derive some knowledge of the patterns of the geographical distribution of components of the energy balance of the Earth's surface from data in the *Atlas of the Earth's Heat Balance* (1963). The atlas contains world maps of average yearly and monthly magnitudes of total radiation, radiation balance of the Earth's surface, heat expenditures on evaporation, and turbulent flows of heat and evaporation. It also includes maps of yearly flows of heat from the surface of oceans to lower layers of water, the radiation balance of the Earth-atmosphere system, the inflow of heat from the condensation of water vapour in the atmosphere, and the transfer of heat by horizontal movements within the atmosphere.

In making use of the maps in the atlas, let us first pause to consider the geographic distribution of total solar radiation, which is the principal component of the radiation balance. The geographical distribution of total radiation is shown on the maps in Figures 4, 5 and 6.

As in all other maps in the atlas, mountain areas that are not dealt with in literature are represented by shaded areas.

It may be seen from Figure 4 that the yearly magnitudes of total radiation change substantially within a range of less than 60 kcal/cm^2·year and values that exceed 220 kcal/ /cm^2·year. At high and middle latitudes the distribution of total radiation is largely zonal in character, while in tropical latitudes we find substantial deviations from zonal distributions. In low latitudes a perceptible reduction in the total radiation may be seen near the Equator, which is associated with increased cloudiness.

Total radiation is greatest in high-pressure belts in the Northern and Southern hemispheres, especially in the desert areas of continents. The highest value for total radiation occurs in Northeast Africa and is attributable to the very insignificant cloudiness over that region.

The reduction in total radiation takes place in regions of monsoon climates (for example, along the eastern coast of Asia) and in certain other regions in which there is increased cloudiness.

In order to describe the distribution of total radiation within a year the maps in Figures 5 and 6 indicate total radiation for December and June, i.e., for those months in which the average height of the Sun is lowest and greatest respectively in the corresponding hemispheres.

On the map representing the total radiation in December, the zero isoline for all longitudes passes somewhat to the

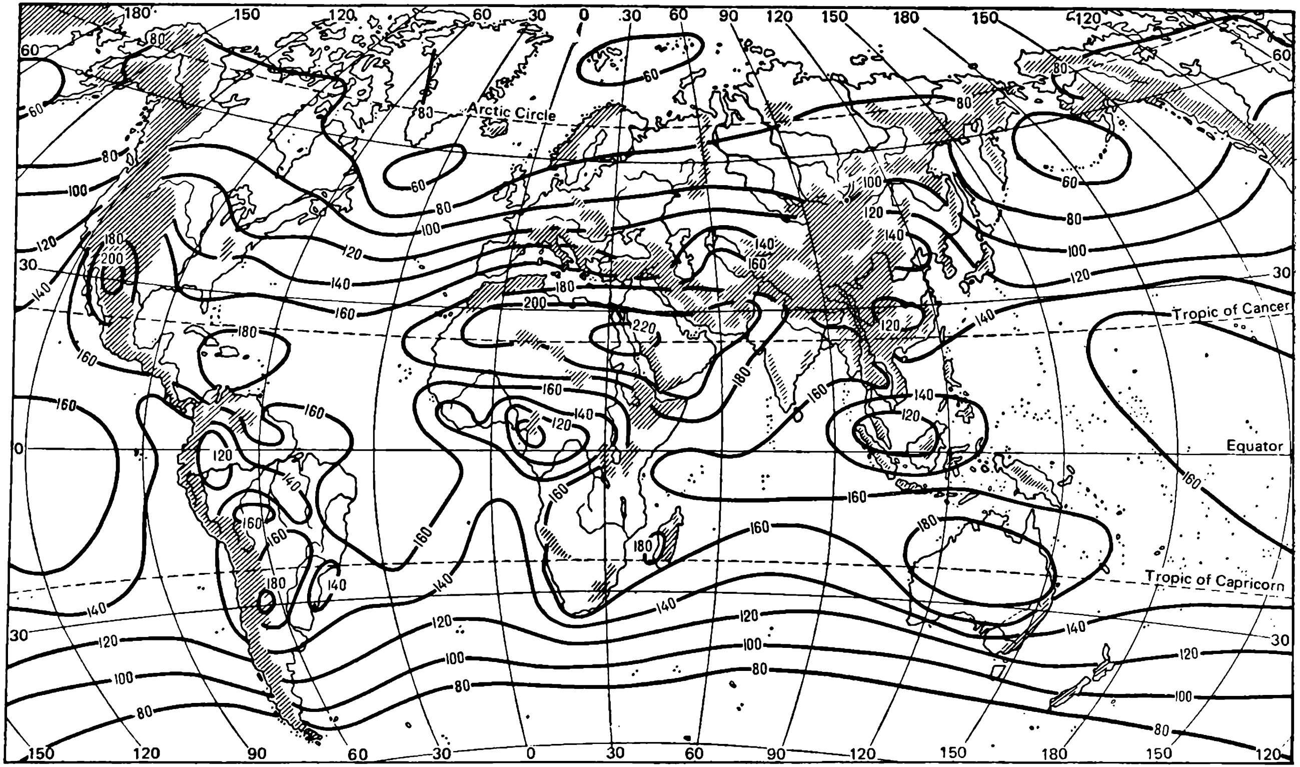


Fig. 4. Total Solar Radiation (kcal/cm²·year) over a Year

north of the Arctic Circle. At higher latitudes during that month the Sun does not appear over the horizon and total radiation is zero.

To the south of the zero isoline total radiation increases rapidly and its distribution outside tropical latitudes of the Northern Hemisphere is largely of a zonal character. At low latitudes however the zonal principle vanishes, and regions of high and reduced radiation are distributed in accordance with the location of regions of greater and smaller cloudiness. At the same time for conditions prevailing in December average zonal changes in total radiation turn out to be relatively small beginning with low latitudes in the Northern Hemisphere and throughout the entire Southern Hemisphere. Negligible reductions in total radiation in the Southern Hemisphere are explained by increases in the length of the day which compensate for a reduction in the average height of the Sun as the latitude increases.

A similar situation exists with regard to the distribution of total radiation in June (Fig. 6). In the Northern Hemisphere the total radiation changes relatively little (even though regions of increased values in desert areas are noticeable), while at middle and high latitudes of the Southern Hemisphere total radiation declines rapidly as the latitude increases.

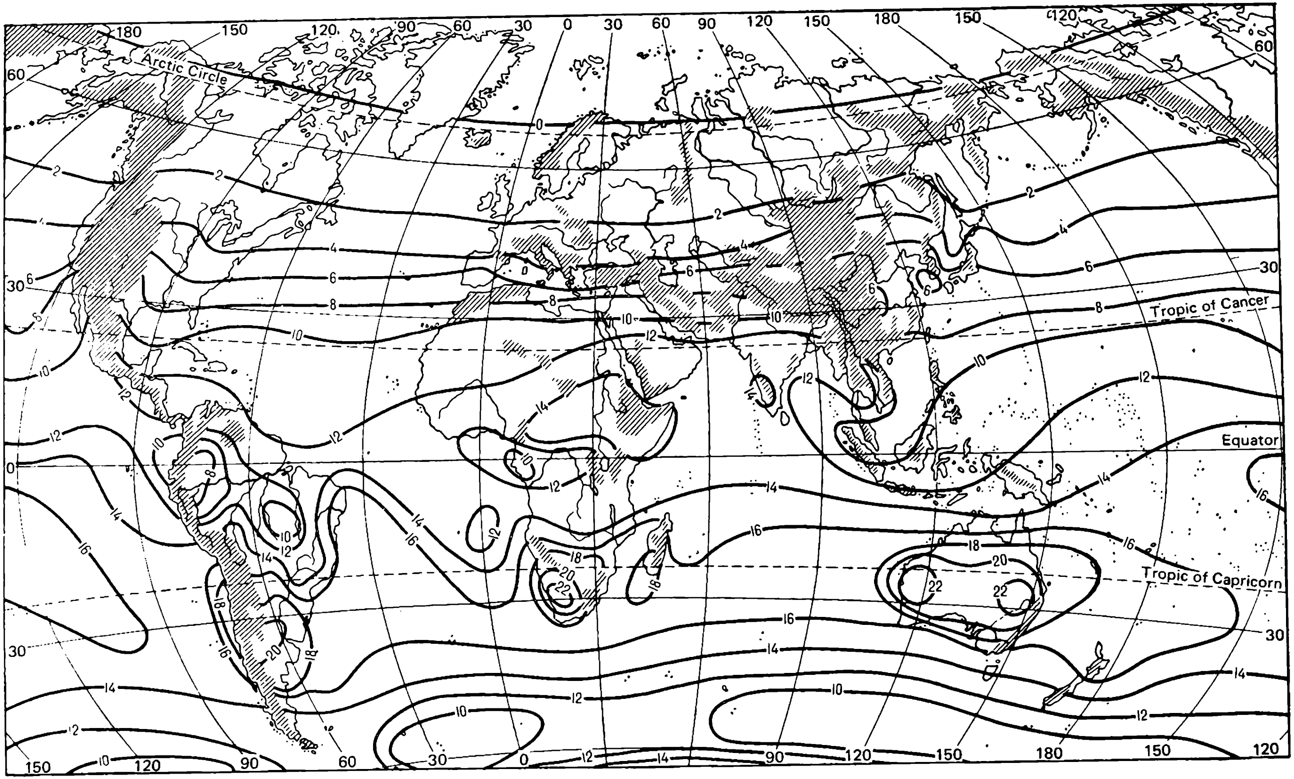
The spatial distribution of the radiation balance at the Earth's surface is shown in Figure 7.

On the map representing the yearly total radiation balance (Fig. 7) the sharp changes in the radiation balances as one passes from land to the sea are clearly visible and are represented by breaks in the corresponding isolines. As the values for the albedo of ocean surfaces are lower, their radiation balance is usually greater than that of land areas at the same latitudes.

It is important to note that yearly total radiation balance at the Earth's surface is positive over the entire surface of oceans and land. Recent studies have shown that negative yearly sums of radiation balance may apparently be found only in regions with permanent snow or ice covers.

Figure 7 shows that the distribution of radiation balance on the surface of oceans is generally zonal in character, although certain deviations from zonality exist, in particular in regions where warm and cold currents operate.

One should also note the relatively small magnitude of


180
150
120
90
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30
0
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60
90
120
150
180
150
120
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Arctic Circle
Tropic of Cancer
Equator
Tropic of Capricorn
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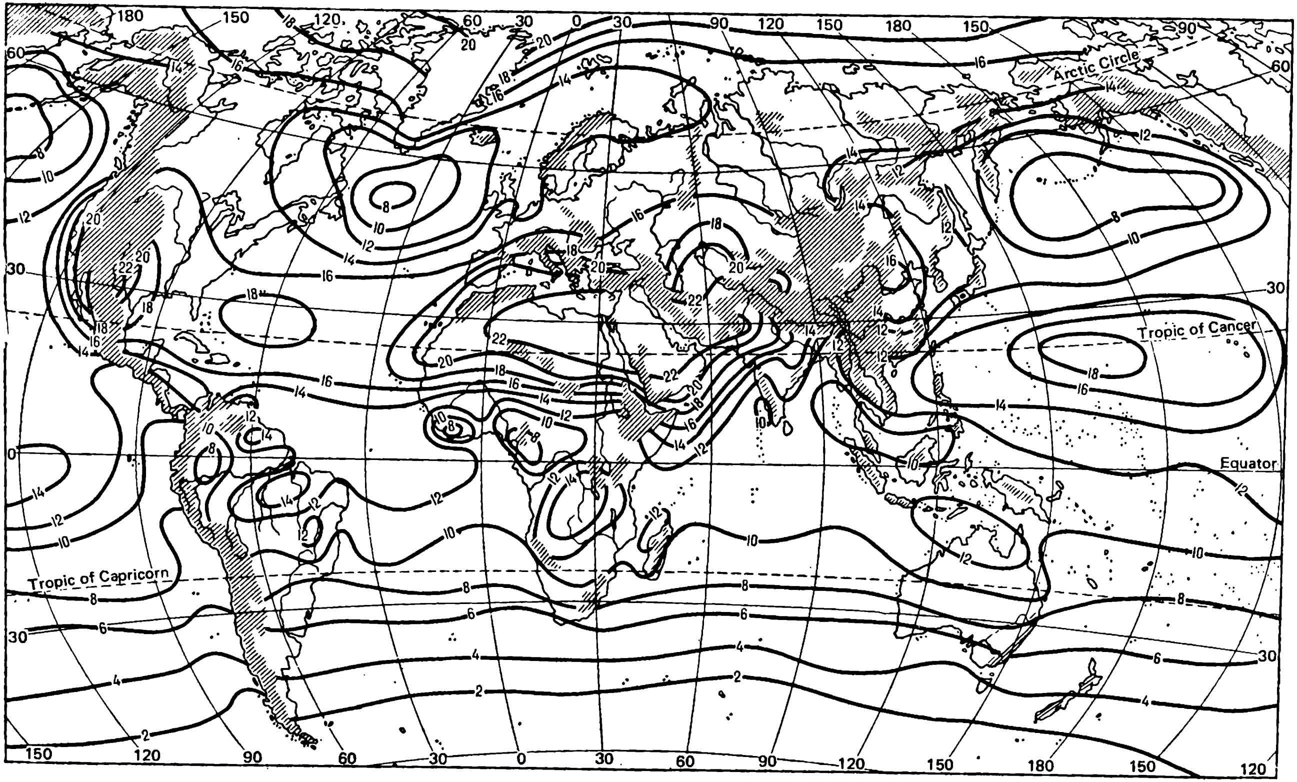


Fig. 6. Total Solar Radiation (kcal/cm²·month). June

changes in the radiation balance of ocean surfaces at tropical latitudes and the rapid reduction in corresponding balances at middle latitudes from lower to higher ones.

The greatest value of the radiation balance occurs to the north of the Arabian Sea where it is more than 140 kcal/cm^2·year.

Changes in the yearly values of the radiation balance on land are also partly zonal in character, but in a number of regions this principle is sharply disrupted through the influence of differences in moisture conditions. One should note the declines in the radiation balance in drought regions by comparison with regions of sufficient or excessive moisture at the same latitudes. This is explained by significant increases in the expenditures of radiation heat on effective radiation in dry climate (as a result of high temperature at the surface, small cloudiness, and the relatively low air humidity) and on the reflection of shortwave radiation.

One can therefore see in Figure 7 that together with a general reduction in the radiation balance at higher latitudes there are visible regions of reduced radiation balance that are attributable to dry climates. In particular this principle may be seen to operate in the Sahara, in the deserts of Central Asia, and in a number of other desert and arid regions. In monsoon regions yearly values of the radiation balance of the Earth's surface are also somewhat reduced through intense cloudiness during the warm season.

It is in humid tropical regions that one finds the highest yearly values of the radiation balance on land, but even there they barely reach 100 kcal/cm^2·year, which is less than the corresponding maximum value for the ocean.

Let us turn our attention to the distribution of other components of the thermal balance of the Earth's surface: heat expenditure on evaporation, turbulent heat exchange, and the redistribution of heat through currents.

The distribution of yearly values of heat expenditure on evaporation is represented in Figure 8. That map shows that the magnitudes of evaporation from land surface and oceans in the vicinity of coastlines differ substantially. This may apparently be explained both by differences in the value of possible evaporation on land and on oceans, and by the influence of insufficient moisture in many land areas, which limits the intensity of evaporation processes and of heat expenditure on evaporation.

It may be seen from the map that at extratropical latitudes heat expenditure on evaporation generally decreases with increasing latitudes in terms of their absolute value. This pattern is altered by major nonzonal changes for both oceans and land. In tropical latitudes the distribution of heat expenditure on evaporation is also quite complex in nature, and compared with high-pressure regions the value of that component of the heat balance declines somewhat in ocean regions adjoining the Equator.

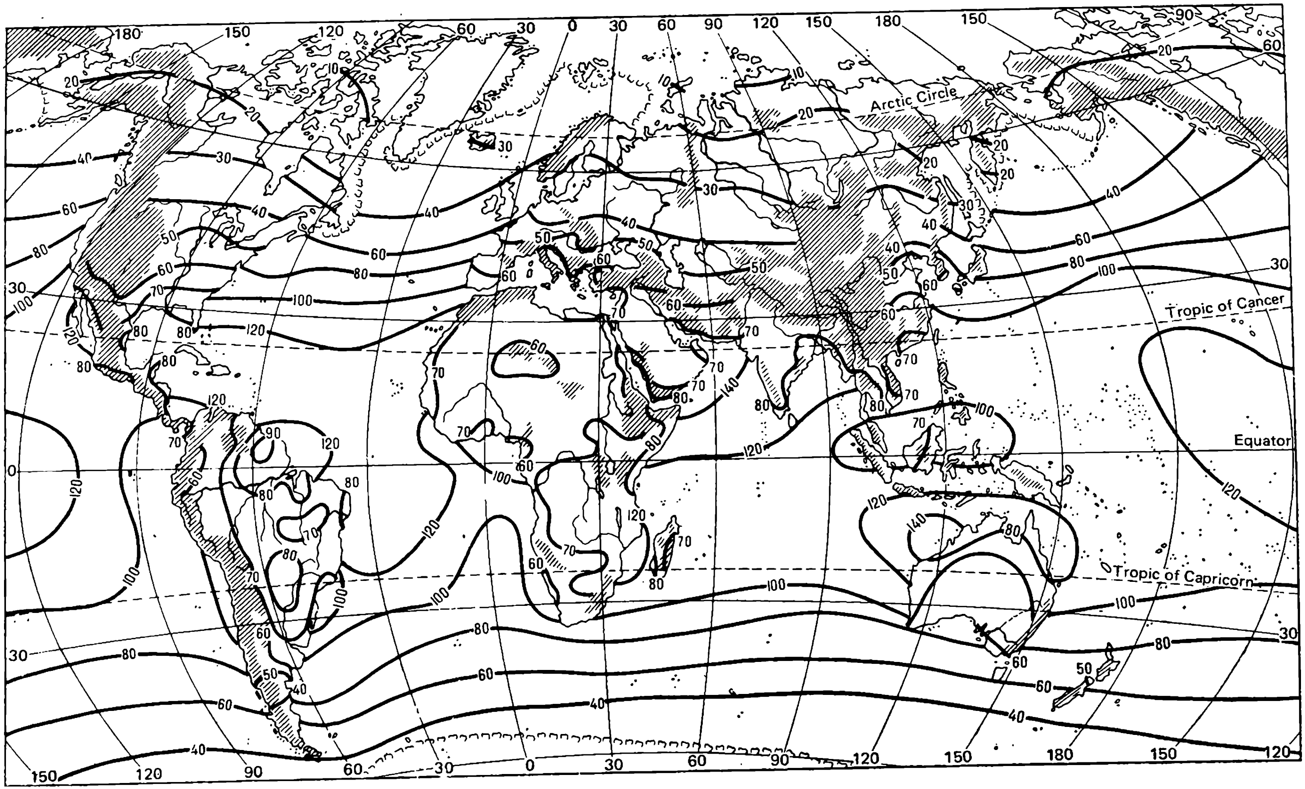
In the case of oceans the principal cause of nonzonal changes in heat expenditure on evaporation is the distribution of warm and cold currents. All major warm currents increase heat expenditure substantially while cold ones reduce it. This may be clearly seen in regions influenced by such warm currents as the Gulf Stream and the Kuroshio, and also by such cold currents as those of the Canary Islands, Bengal, California, Peru, and Labrador. Depending on whether currents increase or reduce water temperature, the magnitude of yearly evaporation from the ocean surface at a given latitude may change by several times.

Aside from ocean currents, nonzonal changes in the values of evaporation and heat expenditure on evaporation from oceans are influenced by conditions of atmospheric circulation determining wind velocity and the air humidity deficit over the oceans.

The distribution of heat expenditure on evaporation from land surfaces deviates from zonal patterns even more. This is because of the very great influence of climatic moisture conditions on evaporation.

When soil moisture is sufficient, evaporation and heat expenditure on evaporation are governed largely by the radiation balance. Such conditions exist at high latitudes and in humid regions at middle and tropical latitudes.

In regions of insufficient moisture the magnitude of evaporation is reduced due to the insufficiency of soil moisture, while in the desert and semi-desert areas it approaches the total yearly precipitation in these regions (which is low). Regions of sharply reduced heat expenditure on evaporation in the Earth's major arid regions are clearly seen in Figure 8. The highest heat expenditure on evaporation is found in certain equatorial regions, where, in the case of abundant


Arctic Circle
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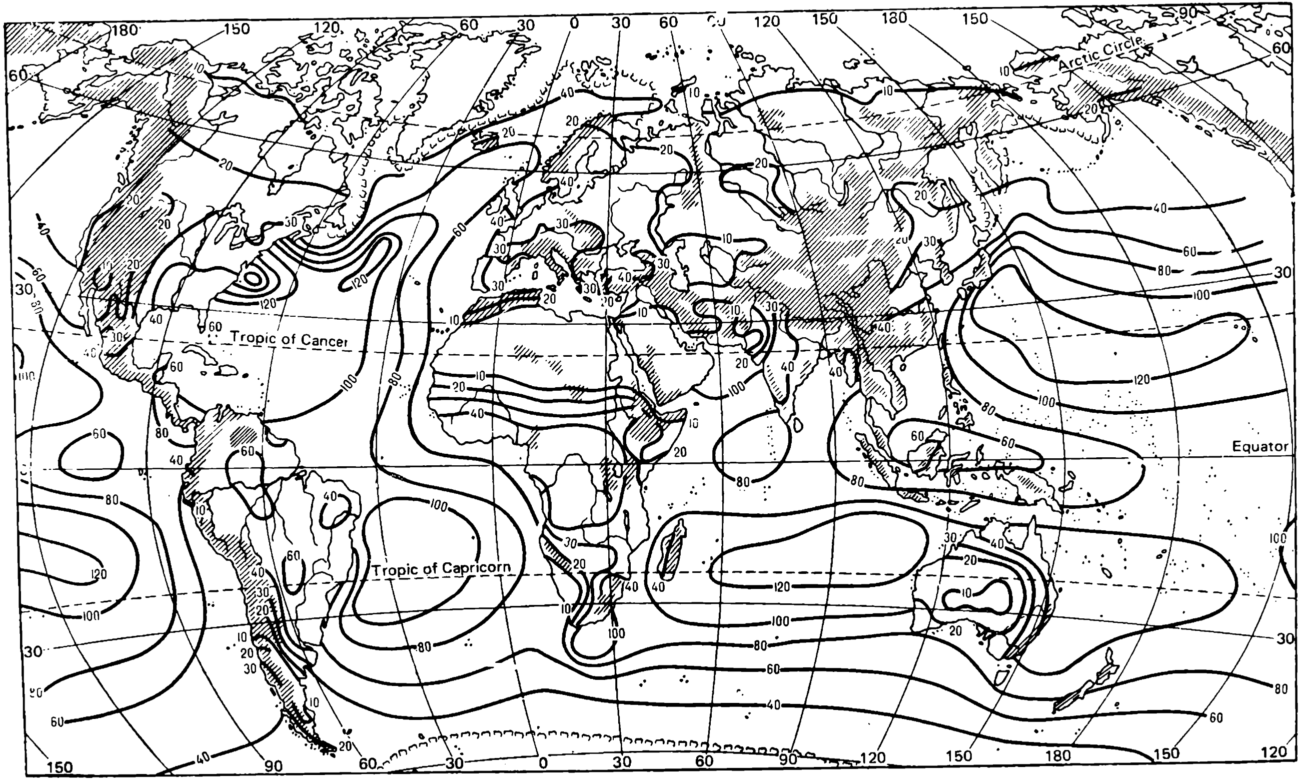

Fig. 8. Heat Expenditure on Evaporation (kcal/cm²·year) over a Year

moisture and large inflows of heat, it exceeds 60 kcal/ /cm²·year. This corresponds to the evaporation over a year of a layer of water more than one meter thick.

Evaporation from ocean surfaces may be much larger, for in such cases the radiation balance is somewhat higher and, what is particularly important, evaporating surfaces may receive a large additional quantity of heat energy through the redistribution of heat by ocean currents. As a result seasonal evaporation from ocean surfaces in a number of tropical and subtropical areas exceeds two meters.

The maps representing monthly heat expenditure on evaporation display the different patterns of seasonal evaporation in extratropical latitudes on land and on oceans.

Evaporation on land declines substantially during the cold season, and depending on moisture conditions attains a maximum either at the beginning or in the middle of the warm season. In contrast to this in the case of oceans evaporation usually increases in cold seasons. The direct cause is a greater difference in temperatures of the water and the air at that time of the year which increases the difference in concentrations of water vapour on the surface of the water and in the air. In addition in many oceanic regions average wind velocities are greater in cold seasons and this, too, serves to increase evaporation.

The map in Figure 9 shows the quantity of heat that is released by the Earth's surface into the air (positive values) or else is received from the air (negative values) during the year. Let us note that over a year the surfaces of all continents (Antarctica excluded) and the larger part of ocean surfaces on the average release heat into the atmosphere.

By comparison with the principal members of the heat balance equation, the magnitude of the turbulent heat flow is not large for the greater part of ocean surfaces, and usually does not exceed 10-20 per cent of their magnitudes. It is in regions in which the water is on the average much warmer than the air, i.e., in regions where powerful warm currents are active (such as the Gulf Stream) and at high latitudes where the sea is still free from ice, that one finds large absolute values of the turbulent flow of heat. Under such conditions the turbulent flow of heat may exceed 30-40 kcal/cm²·year.

Cold currents which reduce the water's temperature also reduce turbulent flows of heat from the surface of the ocean

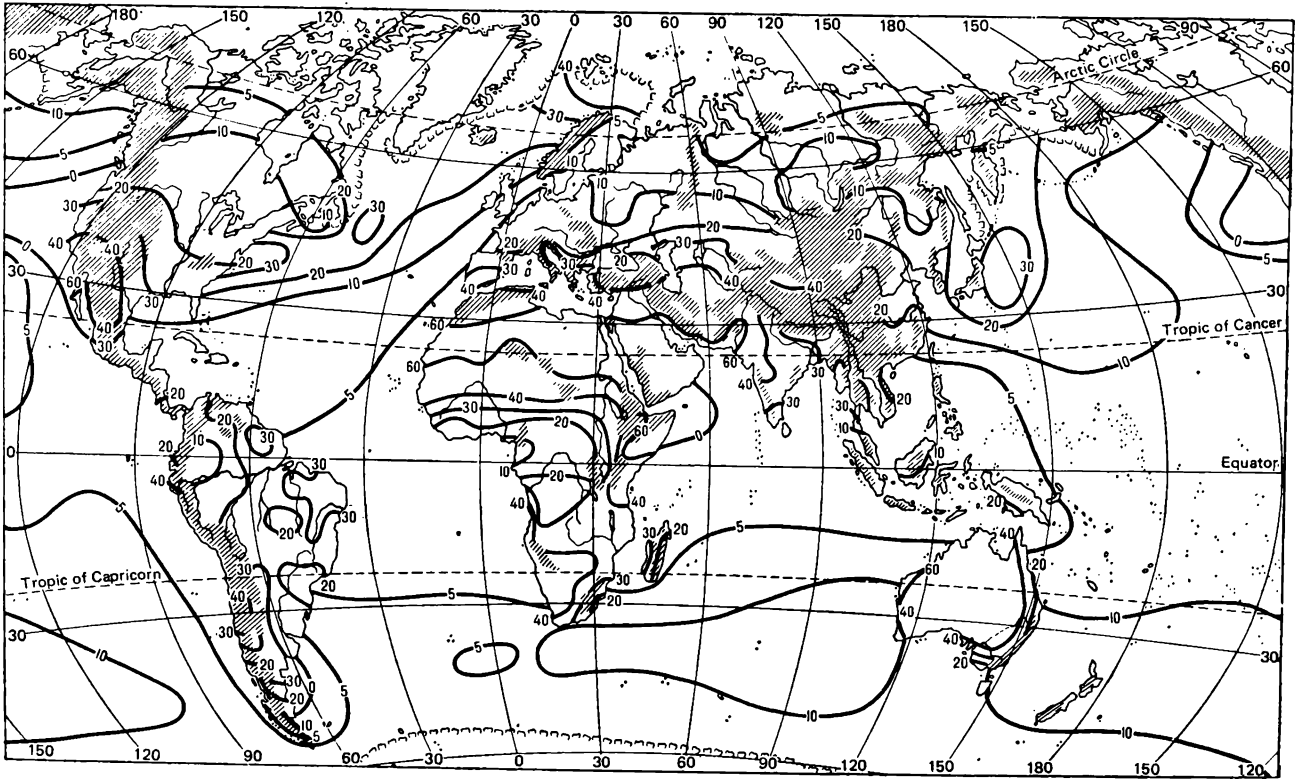

Fig. 9. Turbulent Heat Flows from the Earth's Surface to the Atmosphere (kcal/cm²·year) over a Year

into atmosphere and increase flows in the reverse direction.

Unlike in oceans, where the absolute value of the turbulent flow of heat increases on the average as one moves to higher latitudes, on land turbulent flows of heat change in the opposite direction. At the same time the magnitude of turbulent flows on continents is greatly influenced by climatic moisture conditions: in arid regions the turbulent release of heat from land surface into the atmosphere is much higher than it is in humid regions.

The highest expenditure of turbulent heat flows on land is found in tropical deserts, where it may exceed 60 kcal/cm^2•year. In humid tropical regions, and especially in regions at middle and high latitudes, heat expenditure through turbulent flows is usually much lower.

In the heat balance of oceans inflow or expenditure of heat resulting from horizontal heat exchanges (i.e., primarily as a result of the influence of ocean currents) is very important. Its average yearly values are shown on the map in Figure 10.

Above all the map shows that a large quantity of heat is redistributed in the oceans between tropical and extratropical latitudes. Both warm and cold currents play a very large role in this redistribution, and one can see a correspondence between regions of increased positive values of that particular component of the heat balance (this reflects the outflow of heat from ocean surface to lower layers) and regions with cold currents, and between regions of reduced negative values and regions with warm currents.

In particular such a correspondence may be seen for both major warm currents (the Gulf Stream, the Kuroshio, and the Southwest Pacific Stream) and cold ones (Canary Islands, Bengal, California, and Peru).

L. A. Strokina (1963, 1969) has studied changes in the heat content of the oceans' upper layers over a year. These studies contain average monthly values for such changes in heat content throughout the entire World Ocean together with the corresponding maps. The maps show that changes in the heat content of upper layers of oceanic water at all latitudes may in fact attain significant magnitudes that are comparable with those of the heat balance's main components. It is in the Northwestern regions of the Pacific Ocean and in the adjoining seas that one finds the greatest changes in heat content (over 25 kcal/cm^2·month).

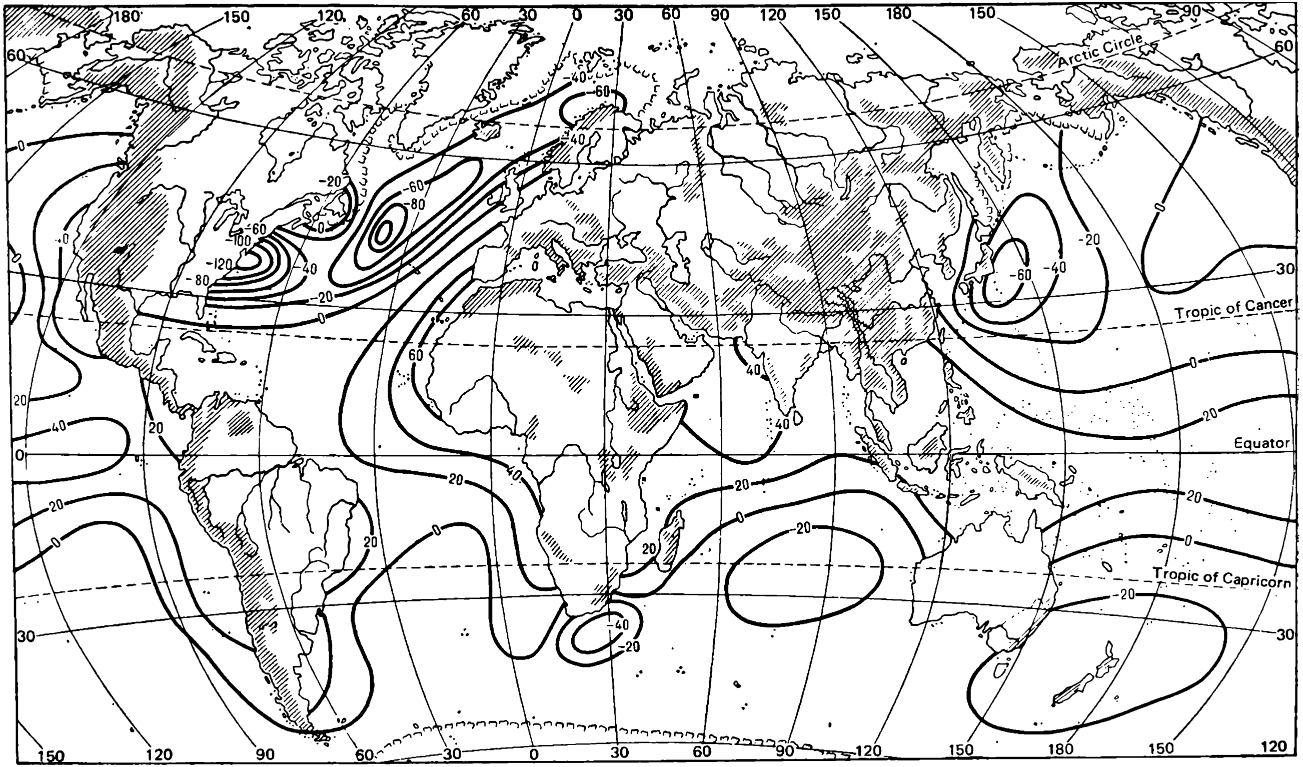

Fig. 10. Heat Flows from Ocean Surface to Underlying Layers of Water (kcal/cm²·year) over a Year

The Earth's Energy Balance. The average values at different latitudes of the heat balance's major components at the surface of continents, oceans, and of the Earth as a whole have been calculated from the data on spatial distribution of these components (Table 1).

Table 1

AVERAGE VALUES OF THE EARTH'S SURFACE ENERGY BALANCE COMPONENTS AT VARIOUS LATITUDES

(kcal/cm²·year)

Latitude	Oceans				Land			Earth			
	R	*LE*	*P*	*A*	*R*	*LE*	*P*	*R*	*LE*	*P*	*A*
70°-60° North	23	33	16	—26	20	14	6	21	20	9	—8
60°-50° »	29	39	16	—26	30	19	11	30	28	13	—11
50°-40° »	51	53	14	—16	45	24	21	48	38	17	—7
40°-30° »	83	86	13	—16	60	23	37	73	59	23	—9
30°-20° »	113	105	9	—1	69	20	49	96	73	24	—1
20°-10° »	119	99	6	14	71	29	42	106	81	15	10
10°-0° »	115	80	4	31	72	48	24	105	72	9	24
0°-10° South	115	84	4	27	72	50	22	105	76	8	21
10°-20° »	113	104	5	4	73	41	32	104	90	11	3
20°-30° »	101	100	7	—6	70	28	42	94	83	15	—4
30°-40° »	82	80	9	—7	62	28	34	80	74	12	—6
40°-50° »	57	55	9	—7	41	21	20	56	53	9	—6
50°-60° »	28	31	8	—11	31	20	11	28	31	8	—11
For the Earth as a whole	82	74	8	0	49	25	24	72	60	12	0

The data in that table confirm that the radiation balance distribution along latitudes is similar on land and in the oceans. In both cases maximum values are observed in the tropics, average values being nearly constant at tropical latitudes.

On land the mean latitudinal value of heat expenditure on evaporation attains its principal maximum at the Equator. These values diminish within the subtropical high-pressure belt. As the latitude increases (in both hemispheres) a certain increase in evaporation occurs (it is more pronounced in the Northern Hemisphere) that is attributable to increased precipitation as compared with arid zones at lower latitudes.

In the oceans, on the other hand, maximum mean latitudinal heat expenditure occurs within high-pressure belts. At latitudes 50-70°, moreover, where radiation balances of land and of oceans are approximately the same, heat expenditure on evaporation is substantially larger for oceans. This is evidently due to large expenditure of heat brought by ocean currents.

Turbulent heat exchange in the oceans gradually increase at higher latitudes. On land it attains a maximum value within high-pressure belts, declines somewhat near the Equator, and drops sharply at high latitudes. These very different patterns of change on land and in the oceans reflect differences in the mechanism of air mass transformation on the surface of continents and of oceans.

Mean latitudinal values of inflow and expenditure of heat in oceans under the influence of currents show that ocean currents carry heat away mainly from a zone ranging from 20° North latitude to 20° South latitude, and that the maximum of the heat absorbed by currents is slightly shifted to the north of the Equator; that heat is carried to higher latitudes and is largely expended in the region of 50°-70° North latitude where warm currents are especially strong.

Mean latitudinal distribution of the components of the energy balance for the Earth as a whole is sometimes characterized by patterns that are typical for continents and sometimes by others that are typical for oceans, depending on the relative proportions of the areas of each in particular zones.

The average values of energy balance components for individual continents and oceans are shown in Table 2.

These data show that in three continents out of six (Europe, North America and South America) the greater share of energy of the radiation balance is expended on evaporation. The opposite is true of the remaining three continents (Asia, Africa and Australia) where dry climates prevail.

The energy balance components of the three oceans differ little from each other, and for each ocean the sum of heat expenditure on evaporation and turbulent heat exchange is close to the magnitude of the radiation balance. This means that the heat exchange among different oceans resulting from currents does not exert any substantial influence on the heat balances of individual oceans.

The values of the components of the heat balance for the Earth as a whole are presented in the last row of Table 1.

These show that in the oceans approximately 90 per cent of the heat radiation balance is expended on evaporation, and only 10 per cent on the direct turbulent heating of the atmosphere. These magnitudes are nearly the same on land. For the Earth as a whole 83 per cent of the radiation balance heat is expended on evaporation, and 17 per cent on turbulent heat exchange.

Table 2

ENERGY BALANCE OF CONTINENTS AND OCEANS
($kcal/cm^2 \cdot year$)

Continents	*R*	*LE*	*P*	Oceans	*R*	*LE*	*P*
Europe	39	24	15	Atlantic	82	72	8]
Asia	47	22	25	Pacific	86	78	8
Africa	68	26	42	Indian	85	77	7
North America	40	23	17				
South America	70	45	25				
Australia	70	22	48				

Data on the average values at various latitudes of the components of the energy balance for the Earth-atmosphere system and also for the atmosphere may be derived from the maps in the *Atlas of the Earth's Heat Balance.* In particular the data for average yearly conditions show that the relative proportions of various components of the Earth-atmosphere energy balance change perceptibly at various latitudes.

In the equatorial zone a substantial inflow of heat produced by changes in the state of water (through condensation and evaporation) is added to the large inflow of radiation energy. These sources of heat produce large expenditure of heat on atmospheric and oceanic advection, for which a relatively narrow zone adjoining the Equator is an extremely important source of energy.

At higher latitudes (up to 30°-40°) a positive radiation balance that declines with increasing latitudes is accompanied by substantial expenditures of energy on water exchange. In most parts of that zone the heat expended on such exchange is comparable to the magnitude of the radiation balance, and the volume of heat redistributed by air and water currents is therefore relatively small.

A zone of the negative radiation balance is found at latitudes above 40°, and its absolute value increases at higher latitudes. The negative radiation balance in that zone is compensated by an inflow of heat that is brought by air and ocean currents. The relative proportions of those components within that balance that compensate for the insufficiency of radiation energy vary at different latitudes. For the belt lying between 40° and 60° the major source of heat is the excess of energy that is released through the condensation of water vapour over the energy that is expended on evaporation from the Earth's surface. At these latitudes the inflow of heat redistributed by ocean currents also plays a large role. The redistribution of heat through atmospheric circulation serves as a major source of heat energy at higher latitudes, and especially in polar regions, where the inflow of heat from condensation is modest, and the influence of ocean currents is either altogether absent (as in the South polar zone) or else is weakened by a permanent ice cover (in the North polar zone).

Calculations show that the atmospheric average radiation balance at various latitudes changes less than do other components of the heat balance. The large absolute negative values for the atmospheric radiation balance that are observed at all latitudes are largely compensated by inflows of heat from condensation.

The role of heat arriving from the Earth's surface as a result of turbulent heat exchange is less important, even though its influence on the atmospheric heat balance is perceptible.

The average values of various components of the energy balance of the Earth-atmosphere system over six-month periods at various latitudes are given in Table 3. Within it Q_a denotes the magnitude of the radiation absorbed by the Earth-atmosphere system, and I_s is the magnitude of the outgoing long-wave radiation at the top of atmosphere.

The upper half of the left side and the lower half of the right side of the table describe warm six-month periods, while the lower half of the left part and the upper half of the right part refer to cold six-month periods.

Table 3

MEAN LATITUDINAL VALUES OF ENERGY BALANCE COMPONENTS OF THE EARTH-ATMOSPHERE SYSTEM FOR SIX-MONTH PERIODS
(kcal/cm²·month)

Latitude	First Six-Month Period (4-9)					Second Six-Month Period (10-3)				
	Q_a	F_a	F_o	B_s	I_s	Q_a	F_a	F_o	B_s	I_s
80°-90° North	7.8	−4.5	0	0.8	11.5	0.1	−9.1	0	−0.8	10.0
70°-80° »	8.2	−4.4	0	0.8	11.8	0.5	−9.3	0	−0.8	10.6
60°-70° »	11.5	−1.8	−0.4	1.2	12.5	1.8	−6.7	−1.5	−1.2	11.2
50°-60° »	14.6	0.0	−1.3	2.8	13.1	4.0	−4.7	−0.5	−2.8	12.0
40°-50° »	16.9	1.0	−2.0	3.9	14.0	6.5	−2.9	0.6	−3.9	12.7
30°-40° »	19.2	1.4	−1.7	4.3	15.2	9.5	−0.9	0.7	−4.3	14.0
20°-30° »	20.0	2.0	−0.4	2.9	15.5	13.7	0.9	0.6	−2.9	15.1
10°-20° »	19.7	2.4	0.9	1.4	15.0	17.0	1.8	1.0	−1.4	15.6
0°-10° »	18.4	1.4	2.2	−0.1	14.9	18.7	2.1	1.4	0.1	15.1
0°-10° South	18.0	2.3	2.2	−1.5	15.0	19.7	2.4	0.8	1.5	15.0
10°-20° »	16.2	2.6	0.9	−2.5	15.2	20.7	3.5	−0.3	2.5	15.0
20°-30° »	13.0	1.4	0.3	−3.4	14.7	20.6	3.5	−1.2	3.4	14.9
30°-40° »	8.9	−0.4	−0.4	−4.2	13.9	18.9	2.0	−1.4	4.2	14.1
40°-50° »	5.9	−1.9	−1.6	−3.5	12.9	15.9	0.0	−0.6	3.5	13.0
50°-60° »	3.3	−3.9	−2.6	−2.5	12.3	12.8	−2.6	0.6	2.5	12.3
60°-70° »	1.0	−9.5	0	−0.8	11.3	8.1	−4.3	0	0.8	11.6
70°-80° »	0.2	−9.8	0	0	10.0	4.4	−6.5	0	0	10.9
80°-90° »	0.0	−8.8	0	0	8.8	3.4	−6.9	0	0	10.3

The data in the table show that absorbed radiation is not the only factor determining the magnitude of outgoing radiation. For middle and high latitudes during cold half-year periods (and for high latitudes in the Southern Hemisphere throughout the entire year) the main source of heat is its transfer from lower latitudes through atmospheric circulation.

Of the two heat exchange processes taking place in oceans the seasonal accumulation and expenditure of heat in the oceanic waters is more important. At some latitudes the value of that component of the energy balance attains 25-30 per cent of the outgoing radiation. The role of redistribution by ocean currents is less important even though at some latitudes it does attain 15-20 per cent of the radiation emitted into outer space. It should be noted that the values of these components of the oceans' energy balance are related to the overall area of the relevant zones at various latitudes. In most cases this reduces their values substantially compared with the components of the energy balance that are related to the overall surface of oceans.

The data in Table 3 show that in order to determine the outgoing radiation from other components of the energy balance all components of the energy balance included in the table should be taken into account.

The values of components of the energy balance for the Earth as a whole are shown in Figure 11.

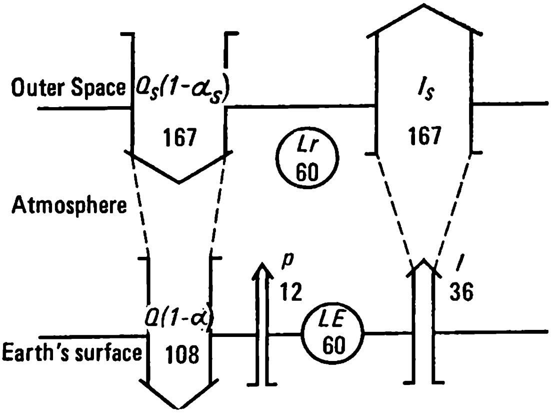

Fig. 11. The Earth's Energy Balance (Components of the Energy Balance in kcal/cm^2·year)

It has already been noted that from an overall flux of solar radiation entering the outer boundary of the troposphere of approximately 1,000 kcal/cm^2·year only one-fourth, or approximately 250 kcal/cm^2·year, passes through a unit surface of the upper boundary of the troposphere because of the Earth's spherical shape.

If we consider that the Earth's albedo is 0.33, we will find that the short-wave radiation absorbed by the Earth is approximately 167 kcal/cm^2·year. On the diagram this magnitude is represented through the arrow $Q_s(1-\alpha_s)$.

A short-wave radiation of 126 kcal/cm^2·year reaches the Earth's surface. The average value of the albedo at the Earth's surface (with due account of differences in the incoming value of solar radiation in various regions) is 0.14. Thus the quantity of solar energy that is absorbed at the Earth's surface is 108 kcal/cm^2·year [the arrow marked $Q(1-\alpha)$], while 18 kcal/cm^2·year is reflected from that surface. It follows from these data that the atmosphere absorbs 59 kcal/cm^2·year, i.e., substantially less than does the Earth's surface.

Since the radiation balance of the Earth's surface is 72 kcal/cm^2·year it follows that the average effective radiation at the level of the Earth's surface is 36 kcal/cm^2·year (the arrow marked I). The overall value of the Earth's long-wave radiation, which is equal to the quantity of absorbed radiation, is close to 167 kcal/cm^2·year (the arrow marked I_s).

It is noteworthy that the ratio of effective radiation from the Earth's surface to the Earth's overall radiation I/I_s is much less than the corresponding ratio of the quantity of absorbed radiation $Q(1-\alpha)/Q_s(1-\alpha_s)$. This difference shows a very large influence of the greenhouse effect on the Earth's thermal processes.

Because of the greenhouse effect the Earth's surface receives approximately 72 kcal/cm^2·year of radiation energy (radiation balance), which is partly expended on the evaporation of water (60 kcal/cm^2·year, represented as the circle marked LE) and partly returned to the atmosphere by turbulent heat losses (12 kcal/cm^2·year, the arrow marked P). The atmosphere's energy balance is therefore composed of the following elements:

1) the inflow of heat from the absorbed short-wave radiation—59 kcal/cm^2·year;
2) the inflow of heat from the condensation of water vapour (in Figure 11 this is represented by the circle Lr)—60 kcal/cm^2·year;

3) the inflow of heat from turbulent heat losses at the Earth's surface—12 kcal/cm^2·year;
4) the expenditure of heat on effective radiation into outer space, which is equal to the difference between the magnitudes of I_s and I, i.e., 131 kcal/cm^2·year.

This last figure corresponds to the sum of the first three components of the heat balance.

III

Climate

1. *Climate and Life*

Climate exerts a decisive influence on living organisms. The geographic distribution of plants and animals, and the nature and intensity of biological processes depend on climatic conditions in many ways. Changes in climate are an important factor in the biosphere's evolution.

Let us consider the principal features of the climate of our own age.

The Present-Day Climate. The climatic conditions of the last century are established on the basis of data provided by instrumental meteorological observation collected through the worldwide network of climatic stations that developed in the second half of the nineteenth century.

Data from these observations show that elements of the meteorological regime change perceptibly over time. Aside from their periodic oscillations (daily and yearly) there are nonperiodic changes in meteorological elements occurring with different time intervals. For short intervals of time (of the order of days or months) nonperiodic changes in the meteorological regime determine oscillations in the weather. These changes are spatially not homogeneous and are largely explained by the instability of atmospheric circulation.

For longer time intervals (of the order of several years), aside from irregular oscillations of individual elements of the meteorological regime, one often finds long-term changes that are similar over large territories. It is such changes that characterize fluctuations in climate.

Since at the present time climatic fluctuations are relatively modest average values of meteorological elements over a period of several decades could be used in order to describe the climatic features of our own age. The use of such averages makes it possible to exclude the influence of unstable atmospheric circulation patterns.

Let us consider a brief survey of observations relating to today's climate giving particular attention to two meteorological elements, namely, air temperature at the Earth's surface and total precipitation reaching the Earth's surface.

Table 4

AIR TEMPERATURES AND PRECIPITATION

	Latitude								
	North								
	90-80	80-70	70-60	60-50	50-40	40-30	30-20	20-10	10-0
Temperature in January (degrees C)	—31	—25	—22	—10	—1	11	19	25	27
Temperature in July (degrees C)	1	2	12	14	20	26	28	28	27
Precipitation (cm/year)	19	26	52	80	75	77	73	114	201

	Latitude								
	South								
	0-10	10-20	20-30	30-40	40-50	50-60	60-70	70-80	80-90
Temperature in January (degrees C)	27	26	25	20	12	5	0	—8	—13
Temperature in July (degrees C)	26	24	18	14	8	1	—12	—30	—42
Precipitation (cm/year)	150	122	91	103	108	101	67	25	11

Table 4 shows average temperature values in January and July at various latitudes as well as average yearly values of atmospheric precipitation at various latitudes.

These show that the difference between average temperatures at the Earth's surface at various latitudes is almost 70°. It is highest at the Equator, where it attains a maximum value, and lowest at the South Pole.

The problem concerning the causes of change in average temperature at various latitudes is examined more closely below. In the present context it should be noted that the Earth's spherical shape exerts a substantial influence on the distribution of these temperatures by producing variations in the total radiation reaching the upper boundary of the atmosphere.

Permanent ice covers are found at high latitudes, where air temperature does not rise above the freezing point throughout the entire year or else most of the year. In the Arctic most of that cover consists of relatively thin ice, while in the Antarctic it consists of a thick continental ice sheet covering nearly its entire surface.

Apart from substantial changes in a meridional direction average air temperature at the Earth's surface also changes substantially in most latitudinal zones at various longitudes. This is largely explained by the distribution of continents and oceans.

The influence of the oceans' thermal regime extends to a large part of the surface of those continents on which a so-called maritime climate exists. This is characterized by a relatively modest yearly oscillations in air temperature at middle and high latitudes. In those extratropical regions of continents in which the influence of the thermal regime of the oceans is less pronounced the amplitude of yearly fluctuations in temperature increase sharply. This corresponds to the characteristics of a continental climate.

The distribution of average latitudinal precipitation values produces a pattern in which the principal maximum value occurs in the equatorial zone, total precipitation declines at subtropical latitudes, two secondary maxima lie at middle latitudes, and precipitation declines as latitudes approach polar regions.

Changes in average values of precipitation at different latitudes are explained by the distribution of average air temperatures and by specific characteristics of atmospheric circulation.

Other conditions being equal, including relative air humidity, total precipitation increases with temperature, since this produces a growth in the volume of water vapour that may undergo condensation. It is evident that if other factors did not exert a substantial influence on precipitation the distribution of its mean latitudinal values would be marked by a single maximum value at low latitudes.

In fact an important role in precipitation is played by vertical air currents that carry water vapour through the condensation level, producing clouds and precipitation.

The atmosphere's overall circulation is closely associated with the geographical distribution of stable pressure systems, among which the low-pressure belt near the Equator is particularly important, as is the high-pressure region at high tropical and subtropical latitudes and also the region of frequent cyclonic formations at middle latitudes.

Since downward movements of air prevail within high-pressure zones this reduces substantially the quantity of precipitation and produces two minima on the curve representing changes in precipitation at different latitudes.

Similarly, the pronounced intensity of upward air motions at equatorial latitudes and in several regions at middle latitudes increases the quantity of precipitation in those areas.

The largest desert areas on Earth are found in the subtropical high-pressure zone, where the quantity of precipitation is negligible. Total precipitation also declines in continental regions distant from the oceans and located at middle latitudes, where the quantity of water vapour carried by air currents from the oceans is small. This reduces relative air humidity and weakens the process of water vapour condensation.

Thus in the continents zones of humid climates are largely located at equatorial latitudes and in regions of maritime climates at middle and high latitudes. Similarly, at high tropical and subtropical latitudes, and in regions of continental climates conditions of insufficient moisture prevail.

Climate Effect on Animate Nature. One of the major factors that limit climatic zones favourable to life is the range of environmental temperature fluctuations within which the life of various plants and animals is possible.

For all plants there exist temperature limits within which their existence is possible. At low temperatures they are harmed as a result of the formation of ice crystals within their tissues. These crystals damage the walls of cells both directly and as a result of the movement of water within such tissues at the moment of its freezing. The magnitude of reductions in temperatures that plants can withstand differs widely for different plants. It also depends on the rate at which temperature is reduced (plants are less vulnerable to slow reduction in temperatures), as well as on the degree of "tempering" acquired by the given plant from its earlier experience in adapting to reductions in temperature.

Similarly, increases in temperature above particular limits also cause harm to plants by destroying their chlorophyll, producing burns on leaves and other forms of damage that may result in the plant's death. An important role in the mechanism through which overheating damages plants is played by the destruction of individual protein compounds within the plant's tissues. The ability of various plants to resist the effects of high temperatures varies widely, and is greatest for succulent plants of deserts.

Laboratory experiments have shown that the rate of photosynthesis and the productivity of autotrophic plants depend on temperature, and that at very high and very low temperatures this productivity generally approaches zero. (That question is examined more closely in Chapter VI.)

Thus thermal factors exert a deep influence on the vital activities of plants. In particular they influence their geographic distribution, their seasonal changes, their composition in terms of species, and the productivity of plant covers.

The influence of temperature on the life activities of animals is equally important. In terms of their heat regime all animals are divided into two groups, namely those whose body temperature changes within a wide range (poikilothermic) and those who maintain their body temperature at an approximately constant level (homoiothermic).

An intermediate group of heterothermic animals is sometimes also mentioned. This refers to animals that are able to regulate their body temperature to a certain extent.

The body temperature of poikilothermic animals usually differs from that of the environment because of the influence of metabolism processes associated with the generation of heat; that heat, however, is usually negligible, owing to which the said difference is, as a rule, relatively small. There are exceptions to this rule that relate to occasions on which such animals are very active. In particular the body temperature of many insects increases substantially when they fly.

Other factors may also cause the temperature of poikilothermic animals to differ from that of the environment. In particular their temperature may decrease through losses of heat produced by evaporation from the animal's body surface and as a result of long-wave radiation into the environment while the absorption of solar radiation may increase body temperature.

In the case of large poikilothermic animals, such as certain reptiles, the influence of thermal inertia within the animal's body may cause fluctuations in body temperature to lag appreciably behind changes in the temperature of the environment.

Homoiothermic animals include the class of birds and that of mammals. In the course of evolution both have acquired effective mechanisms for maintaining their body temperature at a constant level independently of external conditions.

That mechanism is based on the activity of a center of thermoregulation which is located within the brain, namely the hypothalamus. In the presence of changes in external factors governing the animal's heat regime that center provides for corresponding changes in systems of physical and chemical thermoregulation. It is this which permits the animal to maintain a constant body temperature. The principal means for thermoregulation is a wide capacity for change in metabolism, which increases greatly when temperature is reduced. For many animals a large role is also played by changes in evaporation from the surface of their bodies, the state of their hair and other factors.

Wide differences exist in the normal body temperatures of different homoiothermic animals. In the case of birds it is usually 40-42°C, while in mammals it generally lies between 36° and 39°C. The constant character of body temperature of homoiothermic animals is relative, for that temperature frequently changes somewhat in the course of a day as a result of changes in muscular activity, of food ingestion processes and other factors. In external conditions that are anomalous for homoiothermic animals their body temperature may fluctuate perceptibly.

The life activities of both poikilothermic and homoiothermic animals may take place within a certain interval in body temperature that is bounded by so-called lethal temperatures. For a given species of animals these limits may differ substantially depending on their prior existence in high or low temperature conditions.

The lethal temperatures of poikilothermic animals vary considerably. Some animals die as a result of small changes in their body temperature while others are able to adapt to changes in temperature of several tens of degrees.

In the case of homoiothermic animals the difference between the upper and lower lethal temperatures lies most often between 15° and 25°C. For man, whose normal body temperature is approximately 37°C, the lower lethal temperature is 24-25°C, while the upper one is 43-44°C.

The biological functions of both poikilothermic and homoiothermic animals change substantially when changes in body temperature occur. The influence of temperature on life activities is especially pronounced in the case of poikilothermic animals: they are possible only within a specific and sometimes relatively narrow temperature interval. As a result the behaviour of poikilothermic animals is often determined by the need to maintain an optimal body temperature. Thus, for example, in situations in which there is insufficient heat many insects, reptiles and other animals move to sunlit areas and assume positions that produce the greatest possible heating of their bodies from radiation. In cases in which the Earth's surface experiences excessive heat from the Sun, such animals either leave it by burrowing underground, or else moving to branches of plants where the air temperature is lower than that of the Earth's surface. The great influence of thermal conditions on the reproduction of poikilothermic animals should be particularly noted, since this largely determines their ecology in the corresponding periods.

Although the ecology of homoiothermic animals depends on thermal conditions to a lesser extent, in the event of unfavourable conditions these animals also rely on microclimatic features of the landscape in order to bring their body temperatures nearer to optimal ones.

The need for such activities is especially great in the case of small animals whose thermal inertia is negligible. In low environmental temperatures they experience cooling more rapidly because of the large magnitude of the ratio of body surface through which heat is lost to body mass, which determines the heat produced as a result of metabolism (Max Rubner's Law).

The substantial influence of thermal factors on the life activities of animals explains the importance of heat conditions in their environment for their geographical distribution and their ecology.

It is evident, however, that the geographical distribution of various plants and animals depends very much on deviations of elements of meteorological regime from corresponding normals as well as on average climatic conditions.

Data provided by meteorological observations show that at particular moments in time air temperatures at the Earth's surface may deviate from their average monthly value for the given region by several tens of degrees. Large temperature anomalies (especially negative anomalies) often bring about the mass destruction of plants and animals.

The influence of major weather anomalies on animal populations was already noted by Darwin (1859), who described the sharp decline in the number of birds in England following a harsh winter. "The action of climate seems at first to be quite independent of the struggle for existence; but so far as climate chiefly acts in reducing food, it brings on the most severe struggle between the individuals, whether of the same or of distinct species, which subsist on the same kind of food. Even when climate, for example extreme cold, acts directly, it will be the least vigorous individuals, or those which have got least food through the advancing winter, which will suffer most."* It is clear from this observation that large weather anomalies may lead to a mass destruction of animals long before meteorological elements attain values that are lethal for living organisms.

Climatic studies have shown that large anomalies in meteorological elements often extend to large territories stretching over thousands of kilometers. In many cases these regions are wider than the areas that are inhabited by different plants and animals. Large meteorological anomalies may therefore result in the complete extinction of individual species.

Since the magnitude of probable anomalies in air temperature and also in a number of other meteorological elements increases as the time period under consideration becomes longer, it is clear that the influence of fluctuations in climatic conditions on living organisms during the history of the Earth's geological ages may be very considerable. This question is considered more closely in Chapter VII.

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* Ch. Darwin, *The Origin of Species by Means of Natural Selection*, Murray, London, 1859, pp. 68-69.

Aside from being influenced by local thermal regimes the living nature of land is greatly influenced by moisture conditions. Depending on the proportions of precipitation and possible evaporation the soil will contain a specific quantity of water. It has already been noted that soil moisture is one of the major factors governing the life activities of many plants.

The process of photosynthesis can occur only when the stomata in the outer membranes of plants are open. These are the openings that admit carbon dioxide. On land this leads to an unavoidable expenditure of water contained within the plants on transpiration. Together with the expenditure of water on the creation of organic matter and on other physiological functions of plants this must be compensated through the plant's root systems, which extract water from the soil.

The productivity of plant covers is reduced by both excessive water in the soil and water shortages. In particular the process of aeration is inhibited in soils containing large quantities of water. A shortage of oxygen inhibits the development of plant's roots, especially in the case of trees since their roots usually extend to great depths. Under such conditions plants of the marshland type prevail, whose productivity is usually less than that of plants within the same natural zone that are located in areas of optimal moistening.

Insufficient soil moisture also constrains the productivity of plant covers. In such cases the necessary level of transpiration requires that the root systems of plants become very ramified, and this increases the expenditure of organic matter on the respiration of nonproductive parts of plants. At the same time the decline in air humidity that results from low soil moisture reduces the productivity of transpiration (i.e. the ratio of assimilation to transpiration), and this also reduces the quantity of organic matter produced per unit surface occupied by plant covers (cf. Chapter VI).

In cases in which there is a major shortage of soil moisture the plant cover ceases to be continuous, and this results in a further reduction in its productivity per unit area. In semi-desert and desert regions the plant cover consists of individual plants and groups of plants that are distant from each other or else consists of ephemerae, i.e., of plants that exist for a short time when soil moisture is greatest.

In many regions of the Western Hemisphere with dry climates succulent plants occur widely. These are plants possessing a small number of stomata per unit surface. Cacti are among them. Because of the low speed of assimilation that results from their structural peculiarities these plants have a low productivity.

In the driest regions of deserts plant covers are absent altogether.

The dependence of the biomass of animals and of heterotrophic plants on the productivity of autotrophic plants determines the close linkage that exists among all heterotrophic organisms and moisture conditions. In addition, even though the water requirements of heterotrophic organisms on land vary within wide limits, only a small part of these organisms can exist in dry climate, and this also explains the low diversity of living nature in arid regions.

2. The Theory of Climate

The Problems of Climate Theory. Climate theory seeks to identify the average distribution of meteorological elements in both space and time as well as their responsiveness to external climatic factors.

One of the first numerical models in the theory of climate was constructed by M. Milankovich (1920, 1930), who calculated the distribution of average air temperatures at various latitudes on the basis of data on the inflow of radiation at the atmosphere's upper boundary. Milankovich thought that the thermal regime was determined by the radiation heat exchange in each individual latitudinal zone. He did not consider the operation of the atmospheric greenhouse effect or changes in albedo at different latitudes. He also neglected the influences of meridional heat transfers in the atmosphere on the thermal regime as well as of the heat exchange, attributable to changes in the state of water, and the atmosphere's interaction with oceans.

The substantial errors in calculations of temperature fields resulting from a neglect of these factors were compensated through a selection of certain model parameters, which yielded a distribution of average temperatures at various latitudes that did not differ very substantially from observations. That correspondence, however, rested on a relatively arbitrary basis.

In the study of N. Ye. Kochin (1936) a first attempt was made to apply the methods of dynamic meteorology to the formulation of a theory of climate. Kochin calculated the average meridional profiles of pressure and wind velocity in lower layers of the atmosphere on the basis of known distributions of air temperature and pressure at the Earth's surface. Subsequently A. A. Dorodnitsyn, B. I. Izvekov, and M. Ye. Shvets (1939) relied on a similar method in constructing a model of zonal circulation in the Northern Hemisphere during summer months. This line of research activities was further developed in the works of Ye. N. Blinova (1947, etc.) who calculated the average distribution of temperature, pressure and wind velocity in the Northern Hemisphere.

The studies of V. V. Shuleikin (1941, etc.) have greatly influenced subsequent studies in the theory of climate by demonstrating the important and often decisive influence on climate formation of interactions between the ocean and the atmosphere.

Since that time many new models of climate were developed that made it possible to calculate average fields of meteorological elements. Achievements in the development of computer technology have made possible the design of numerical models reproducing nonaveraged fields of meteorological elements. The calculation of such nonaveraged fields for relatively prolonged time intervals has made it possible to calculate average fields describing climatic conditions.

A large number of such studies has been carried out by J. Smagorinsky, Y. Mintz, and other authors (Smagorinsky, 1963; Smagorinsky, Manabe and Holloway, 1965; Manabe and Bryan, 1969; Holloway and Manabe, 1971, etc.; Mintz, 1965, etc.).

In particular the study carried out by S. Manabe and K. Bryan was based on a numerical model of the theory of climate that included the influence of the circulation of ocean waters on climatic conditions. The studies of J. Holloway and S. Manabe produced world maps representing the basic components of the heat and water balances at the Earth's surface on the basis of a theoretical method. These maps proved to be similar to corresponding maps based on empirical methods.

The theoretical models of climate may be employed to calculate changes in climate produced by natural factors as well as by man's activities.

In order to employ numerical models in studying climatic changes such models must then meet several requirements, which are less rigorous in the case of models applied in studies of the current climatic regime.

The first such requirement is that the model should not include empirical data concerning the distribution of individual elements of climate, particularly those that change substantially in the course of changes in climate.

Second, such models must recognize realistically all types of inflow of heat that influence the temperature field appreciably; in particular, the law of conservation of energy must be met.

Third, the model must include major feedback relations among various elements of climate.

Closer attention will now be given to this last requirement.

Feedback Relations. The stability of climate is maintained by the feedback relations among its elements that may be called negative. Such feedback relations contribute to a reduction in anomalies of meteorological elements and to an approximation of the values of these elements towards their climatic normals. An example of a negative feedback relationship is mentioned in Chapter I, namely the dependence of long-wave radiation on temperature at the Earth's surface. As temperature increases so does the intensity of long-wave radiation thus producing a greater expenditure of heat energy and inhibition of further increases in temperature. Another example of a negative feedback relationship concerns the dependence of heat transfer in the atmosphere on the air temperature gradient. Usually the flow of heat in the atmosphere is directed from a zone of higher temperature to a zone of lower temperature. This produces a smoothing process that serves to eliminate differences in temperature distribution.

With regard to changes in climate a major role is played by positive feedback relationships that cause an increase in anomalies of meteorological elements and thus reduce the climate stability.

Positive feedback relationships include the dependence of the absolute air humidity on its temperature. As temperature rises, evaporation usually increases from water surfaces or moist surfaces, and this leads to a comparative constancy in the relative air humidity in most climatic zones (except for dry continental regions). In such conditions the absolute air humidity increases as temperature rises. This is confirmed by numerous empirical observations.

Since long-wave radiation declines as absolute humidity increases, a growth in absolute humidity with the rise in temperature partly compensates the intensification in long-wave radiation that is attributable to increased temperature.

This dependence was studied by S. Manabe and R. Wetherald (1967), who found that in conditions of a constant relative humidity the influence of changes in the solar constant on air temperature at the Earth's surface is almost twice as great as when absolute humidity is stable. This particular feedback relationship must be incorporated in the numerical models of the thermal regime that are employed in studies of climatic changes.

A still greater role in representing patterns of changes in the atmospheric thermal regime is played by another positive feedback relationship, that resulting from the influence of snow and ice covers on the magnitude of the albedo of the Earth's surface.

The influence of the snow cover on climatic conditions was first studied by A. I. Voeikov (1884, etc.), who established that it contributes to a reduction in air temperature over corresponding surfaces. Subsequently Brooks (1950) concluded that because of their high albedo ice covers reduce air temperature substantially and that changes in climate are intensified by the formation or else the melting of ice.

It is possible to estimate the value of the albedo for the Earth's surface and for the Earth-atmosphere system at high latitudes on the basis of data collected by expeditions in the Arctic and Antarctic regions, as well as on the basis of data provided by the satellites. This value may be compared with the magnitude of the albedo in regions that do not possess snow or ice covers.

Available data indicate that during summer months the ice surface albedo in the Central Arctic is approximately 0.70, while in the Antarctic it is approximately 0.80-0.85. Keeping in mind the magnitude of the average albedo of the Earth's surface in regions that are free of ice and snow which does not exceed 0.15, it may be concluded that, other conditions being equal, snow and ice covers reduce the radiation that is absorbed by the Earth's surface by several times.

Similarly snow and ice greatly influence the albedo of the Earth-atmosphere system. According to data obtained through satellites (Raschke et al., 1973) the albedo of the Earth-atmosphere system in the Central Arctic region is equal to 0.55 during the summer, and to approximately 0.60 in the Antarctic. This is approximately twice the value of the estimated albedo for the planet as a whole that is mentioned in that study, which is about 0.30.

Such large differences in values of the albedo must exert a considerable influence on the atmospheric thermal regime.

When reductions in air temperature at the Earth's surface produce snow and ice covers this creates a sharp decline in the absorbed radiation, thus contributing to a further reduction in the Earth's temperature and to further increases in the area under snow and ice. The reverse process may be effected by increases in temperature when this results in the melting of ice and snow.

The inclusion of this feedback relation into a numerical model of the atmospheric thermal regime has shown that it exerts a very substantial influence on the distribution of air temperature at the Earth's surface (Budyko, 1968).

In order to estimate that influence it is possible to point to a simple example which shows how average global air temperature will change if, in the absence of clouds, the Earth's surface should be fully covered by snow and ice (Budyko, 1962). The Earth's albedo increases perceptibly as compared with its present value, and this will influence air temperature. The Earth's "effective" temperature, which corresponds to its long-wave radiation, is proportional to $\sqrt[4]{1-\alpha_s}$, where α_s is the albedo. If the albedo changes from α_s' to α_s'', the absolute value of effective temperature will change according to the relation $\sqrt[4]{\frac{1-\alpha_s''}{1-\alpha_s'}}$. If one considers the Earth's current albedo to be 0.33 and the albedo of a dry snow cover to be 0.80 we find that when the Earth is covered with snow the average effective temperature should decline by approximately 75°C.

There are grounds for believing that the reduction in average air temperature near the Earth's surface will in fact be larger than this magnitude. At the present time the average temperature in lower layers of the atmosphere rises substantially almost everywhere at the Earth's surface as a result of the greenhouse effect that is associated with the absorption of long-wave radiation by water vapour and carbon dioxide in the atmosphere. But at very low temperatures that effect is not significant, and the formation of dense clouds perceptibly altering radiation flows also becomes impossible. Under such conditions the atmosphere becomes more or less transparent both to short-wave and long-wave radiation.

The average temperature of the Earth's surface for a transparent atmosphere is determined by the simple formula $\sqrt[4]{\frac{S_0(1-\alpha_s)}{4\sigma}}$, where S_0 represents the solar constant and σ is Stefan's constant. It follows from that formula that when $\alpha_s=0.80$ the Earth's average temperature will be equal to 186°K, or —87°C.

Thus even if either snow or ice were to cover the entire Earth even for a short period, its average temperature, which is now 15°C, would decline by approximately 100°C. This estimate shows how enormous the influence of the snow cover may be on the thermal regime.

Attempts to estimate the influence of sea polar ices on the Arctic thermal regime were made in a number of studies (Budyko, 1961; Rakipova, 1962; Donn and Show, 1966; Fletcher, 1966 et al.). On the basis of data on the thermal balance in central regions of the Arctic Ocean and of approximate values of proportions derived from a semi-empirical theory of climate, it was established that polar ice reduces average air temperature in the Central Arctic during the summer months by several degrees, and by approximately twenty degrees in the winter. It was concluded that the Arctic Ocean could be free of ice in the present age, but this state would be extremely unstable and it could develop an ice cover as a result of a relatively small change in climate.

Since a permanent ice cover exerts a substantial influence on the atmospheric thermal regime even when it extends to only a small part of the Earth's surface, this must be taken into account in studies of climatic changes.

A Semi-Empirical Model of the Thermal Regime. While studies relating to a general theory of climate have recorded substantial achievements they usually do not fulfil all the requirements mentioned earlier that must be met by a realistic theory of climatic change.

In view of the difficulties attaching to the use of more general theories of climate in studying its changes, a number of studies have attempted to apply semi-empirical models, in which large-scale atmospheric processes are schematized to make it possible to meet the necessary requirements.

One such study is a semi-empirical model of the atmospheric thermal regime (Budyko, 1968) that was based on the following considerations.

In the absence of an atmosphere the average temperature of the Earth's surface will be determined by an equilibrium in radiation, i.e., long-wave radiation at the Earth's surface will be equal to the absorbed radiation. Let us represent this through the relationship

$$\delta\sigma T^4 = \frac{1}{4} S_o (1-\alpha_p), \tag{3.1}$$

where δ is a coefficient that expresses the difference in the properties of the radiating surface from those of a black body, σ is Stefan's constant, T is surface temperature, S_o is the solar constant, and α_p is the Earth's average albedo.

Assuming that $\delta=0.95$, $\sigma=8.14\cdot10^{-11}$ cal/cm²min·deg⁴, and that $S_o=1.95$ cal/cm²·min, we find from (3.1) that for $\alpha_p = 0.33$ the Earth's average temperature is equal to 255°K or −18°C.

Let us take into account that according to observations average air temperature at the Earth's surface is approximately 15°C. Thus for the same value of the albedo the atmosphere increases average air temperature by approximately 33°C. This is associated with the greenhouse effect, i.e., with a greater transparency of the atmosphere to short-wave radiation than to long-wave radiation.

It should be noted that such an estimate is subject to reservations: in the absence of an atmosphere the planet's albedo cannot be equal to its present value, which is taken to be 0.33. Under present conditions the average albedo of the Earth's surface is equal to 0.14. It would seem that before the appearance of the atmosphere the Earth's albedo was less than that magnitude, and perhaps did not differ much from the Moon's albedo, which is equal to 0.07. For such a value of the albedo the average temperature at the Earth's surface would be approximately 3°C.

In order to estimate the influence of the inflow of solar radiation and of the albedo on average temperature at the Earth's surface under real conditions it is necessary to consider the dependence of the long-wave radiation at the atmosphere's outer boundary on temperature distribution. That dependence may be established from both observed and calculated data relating to long-wave radiation.

By comparing these data with different elements of the meteorological regime it was possible to establish that average monthly values of outgoing radiation largely depend on air temperature in the vicinity of the Earth's surface, and also on cloudiness.

That dependence was expressed in terms of the following empirical formula:

$$I_s = a + bT - (a_1 + b_1 T)\, n, \qquad (3.2)$$

where I_s represents the outgoing radiation in kcal/cm^2·month, T represents the temperature in Celsius degrees, n represents cloud intensity as a fraction of one, the values of its corresponding coefficients being $a = 14.0$, $b = 0.14$, $a_1 = 3.0$ and $b_1 = 0.10$.

In order to verify the formulas that link outgoing radiation with meteorological factors one should apply the condition of equality of outgoing radiation for the entire globe to the quantity of absorbed radiation.

$$Q_{sp}(1 - \alpha_{sp}) = I_{sp}, \qquad (3.3)$$

where the values of Q_{sp}, α_{sp} and I_{sp} refer to the planet as a whole.

In accordance with the considerations relating to the influence of snow and ice covers on the thermal regime that were mentioned earlier it may be shown that equations (3.2) and (3.3) have at least two solutions, one of which corresponds to current climatic conditions, while the other refers to a "white Earth" regime, i. e., when the Earth is completely covered with snow and ice.

Let us assume that for low negative temperatures in air layers adjoining the Earth's surface at all latitudes the Earth will be fully covered by snow and ice, and that the global albedo of the Earth-atmosphere system will approximately correspond to the albedo of the Antarctic region, i.e., 0.6-0.7. It follows from (3.2) and (3.3) that in such a case the Earth's average temperature at the surface will range from —47°C (for lower values of the albedo) to —70°C (for higher values). For such values of the average global temperature a complete covering of the Earth by ice is inevitable. This confirms the point of view that has been expressed earlier concerning the multi-valued correspondence of contemporary climatic conditions to external factors forming the climate.

That problem will be considered more closely below.

From (3.2) and (3.3) let us find the formula for the Earth's average temperature.

$$T_p = \frac{1}{b - b_1 n} [Q_{sp} (1 - \alpha_{sp}) - a + a_1 n]. \qquad (3.4)$$

That formula should be applied within the limits of actual changes in average monthly temperatures at the Earth's surface and its accuracy depends on the accuracy of the relationship (3.2). It should be kept in mind that because of the structure of formula (3.4) the accuracy of calculations of temperature for substantial cloudiness (for values of n close to 1) declines perceptibly by comparison with calculations for low or average cloudiness. Although this does reduce the possibility of applying formula (3.4) that formula does make it possible to arrive at certain conclusions concerning the influence of cloudiness on the thermal regime of lower air layers.

Consideration must then be given to the dependence of the albedo on cloudiness. Such a relationship possesses the form

$$\alpha_s = \alpha_{s_n} n + \alpha_{s_0} (1 - n), \qquad (3.5)$$

where α_{s_n} and α_{s_0} represent the albedo of the Earth-atmosphere system for cloudiness equal to unity and without clouds, respectively.

The following relation follows from (3.4) and (3.5):

$$T_p = \frac{1}{b - b_1 n} \{Q_{sp} [1 - \alpha_{s_n p} n - \alpha_{s_0 p} (1 - n)] - a + a_1 n\}. \qquad (3.6)$$

Assuming (in accordance with the methodology employed in the construction of the maps in *The Atlas of the Earth's*

Heat Balance) that

$$\alpha_{sp} = 0.66\alpha + 0.10, \qquad (3.7)$$

where α is the albedo at the Earth's surface, we find that the average magnitude of $\alpha_{s_0 p}$ for the Northern Hemisphere is 0.20. We then obtain from formula (3.5) that for $\alpha_{sp} = 0.33$ and $n = 0.50$ the average value of $\alpha_{s_n p}$ is 0.46.

Inserting these values of $\alpha_{s_0 p}$ and $\alpha_{s_n p}$ into formula (3.6) we find that for an average value of Q_{sp} for the Earth as a whole the influence of cloudiness on temperature is relatively small and may lie within the limits of calculation errors.

A number of conclusions may be inferred from formula (3.4) concerning the influence of climate-forming factors on average temperature at the Earth's surface.

Changes in solar radiation by one per cent change the average temperature (for cloudiness of 0.50 and for the existing Earth's albedo) by approximately 1.5°C. This estimate is nearly twice as large as the estimate of the corresponding influence in the absence of an atmosphere. Thus the atmospheric radiation properties amplify substantially the influence of changes in radiation on the thermal regime of the Earth's surface.

Changes in albedo by 0.01 change the average temperature by 2.3°C. Thus the thermal regime is very sensitive to changes in the albedo if these changes are not determined by cloudiness. As has been noted earlier the influence of cloudiness on the thermal regime that follows from changes in the albedo is largely compensated by the corresponding changes in outgoing long-wave radiation.

In some of the Earth's latitudinal zones the horizontal redistribution of heat in the atmosphere and hydrosphere influences the thermal regime substantially.

Quantitative estimates of the influence of the horizontal redistribution of heat on the atmospheric thermal regime may be based on data relating to components of the heat balance of the Earth-atmosphere system.

The equation representing the energy balance of the Earth-atmosphere system is:

$$Q_s(1 - \alpha_s) - I_s = C + B_s, \qquad (3.8)$$

where C is equal to $F_s + L(E - r)$, i.e., the sum of the inflows of heat attributable to horizontal movements in the atmosphere and the hydrosphere.

For average yearly conditions the term B_s, which describes the accumulation or loss of heat over the given period of time, is equal to zero, and the inflow of heat C is equal to the value of the radiation balance of the Earth-atmosphere system. Since the values in that balance may be established either on the basis of observations or else through calculations it is evident that one may simultaneously find the values of the horizontal redistribution of heat.

It may be assumed that the value of the term C is related to the horizontal distribution of the average temperature of the troposphere. Since air temperature deviations from mean vertical distributions in the troposphere are small compared with geographic changes in temperature, it may be concluded that average air temperature in the troposphere is closely linked to the temperature at the Earth's surface.

There are thus grounds for assuming that a dependence exists between the horizontal transfer of heat and the distribution of temperature near the Earth's surface.

Let us consider the relation between these magnitudes for average yearly conditions in the Northern Hemisphere in various latitudinal zones. In this connection let us calculate the radiation balance in accordance with the formula

$$R_s = Q_s(1-\alpha_s) - I_s. \tag{3.9}$$

The magnitudes of the radiation balance of the Earth-atmosphere system that are calculated in this manner and which are equal to the term C make it possible to calculate the meridional flow of heat in the Earth-atmosphere system.

Since the meridional transfer of heat is effected in the form of a transfer of heat from warmer regions to colder ones, the magnitude of C may be assumed to depend on $T - T_p$, where T is the average temperature at the given latitude and T_p is the average global temperature within the lower air layer.

In order to study this relationship values of C were compared with corresponding magnitudes of $T - T_p$. It was established that a clear relationship exists between them which may be represented in the form of the following empirical formula:

$$Q_s(1-\alpha_s) - I_s = \beta(T - T_p), \tag{3.10}$$

where $\beta = 0.235$ kcal/cm^2 month·deg.

The existence of such a relationship greatly simplifies calculations of the meridional redistribution of heat in a model of the thermal regime.

From (3.2) and (3.10) we find that

$$T = \frac{Q_s(1-\alpha_s)-a+a_1n+\beta T_p}{\beta+b-b_1n}. \tag{3.11}$$

With the help of that formula it is possible to calculate the average yearly temperature at various latitudes.

In order to study the atmospheric thermal regime in various seasons the model that has been presented should be changed in such a way as to consider a number of additional factors.

Such a more general model has been proposed in a study (Budyko, Vasishcheva, 1971) in which the equation for the energy balance of the Earth-atmosphere system is represented as

$$Q_{s_m}(1-\alpha_{s_m})-I_{s_m}=C_m+B_s; \tag{3.12}$$

$$Q_{s_x}(1-\alpha_{s_x})-I_{s_x}=C_x-B_s, \tag{3.13}$$

where B_s represents the inflow or expenditure of heat attributable to the cooling or heating of the Earth-atmosphere system which is determined largely by the ocean's cooling or heating. Values relating to warm and cold half-year periods are represented by the indexes m and x, respectively.

In order to determine the value of B_s, equations representing the energy balance of the oceans were employed, as well as relations that link the components of the energy balance of oceans with water temperature and air temperature.

The solution to these equations yields formulae for establishing the average temperatures of lower air layers at various latitudes in different seasons. Calculations of temperature distribution on the basis of these formulae produce results that accord well with the observed data.

The Single-Valued Character and Stability of the Climate. The model of the atmospheric thermal regime that has been described above may be employed to study the single-valued character and stability of current climate.

Let us assume, in accordance with empirical data for both sea polar ice and continental ice sheets on plains and mountains in extratropical belts, that the average latitudinal boundary of the permanent snow and ice cover corresponds to an average yearly temperature of —10°C. In the light of this condition let us then use the model of the thermal regime, applied to average yearly conditions, and calculate the dependence of the average latitude of the boundary of the polar ice cover in the Northern Hemisphere on the inflow of radiation energy to the atmosphere's outer boundary.

This dependence is represented in Figure 12, where $\frac{\Delta Q_{sp}}{Q_{sp}}$ represents the relative change in the inflow of solar radiation to the atmosphere's outer boundary expressed in per cent, and φ is the average latitude of the boundary of the polar ice cover in the Northern Hemisphere.

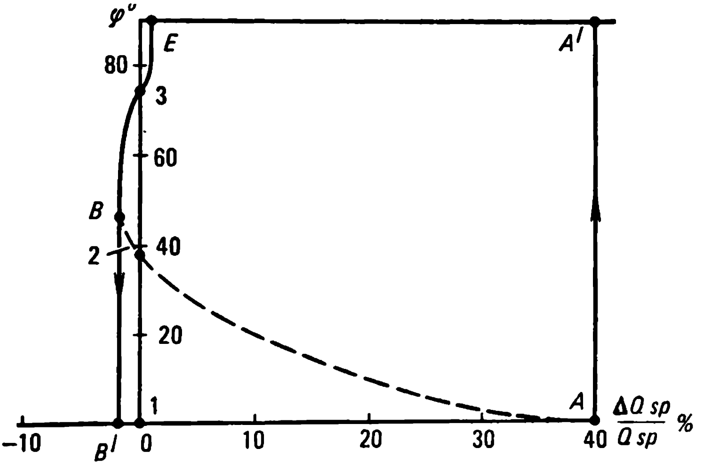

Fig. 12. Dependence of the Average Latitude of the Polar Ice Cover Boundary in the Northern Hemisphere on the Inflow of Radiation to the Atmosphere's Outer Boundary

The dependence of φ on $\frac{\Delta Q_{sp}}{Q_{sp}}$ is represented by the thick lines which show that this relation has an ambiguous character and that it differs substantially in cases in which the inflow of heat to the atmosphere's outer boundary increases or declines.

Let us first consider the case in which the inflow of heat increases by comparison with an initially small magnitude. For small inflows of heat there occurs a full glaciation of the Earth ($\varphi = 0°$), which remains as the inflow of heat increases to its current magnitude (point 1) and to magnitudes exceeding that value by several tens per cent (point A).

The glaciation regime corresponding to point A is unstable and a small increase in the inflow of heat causes it to pass to a regime in which there is no ice whatever (point A'). Further increases in the inflow of heat produce a continuation of the ice-free regime.

A reduction in the inflow of heat from an initial value, substantially exceeding the solar constant, first produces conditions of an ice-free regime ($\varphi = 90°$). After point E is reached, which is close to the current value of the inflow of heat, a polar glaciation develops that subsequently advances rapidly as the inflow of heat declines. After reaching point 3, which corresponds to the current climatic regime, reductions in the inflow of heat by approximately two per cent from the currently observed value cause the ice cover to reach a position 50° North latitude (point B).

The glacial regime corresponding to that point is not stable, and small reductions in the inflow of heat cause it to pass to the regime of full glaciation (point B^1), which then continues with further reductions in the inflow of heat.

The model mentioned earlier also makes it possible to establish the relation between values of φ and $\frac{\Delta Q_{sp}}{Q_{sp}}$ which is represented by the broken line AB. Since that curve represents increases in values of φ as $\frac{\Delta Q_{sp}}{Q_{sp}}$ declines (or else reductions in φ as $\frac{\Delta Q_{sp}}{Q_{sp}}$ increases) it seems likely that it corresponds to an unstable regime of glaciation that passes to regimes of full glaciation or else of a complete absence of ice in response to small fluctuations in the inflow of heat. In such a context the point on line AB that corresponds to the current value of the inflow of heat (point 2) describes an unstable regime that cannot last for a prolonged period of time.

Thus the dependence of polar glaciation on the inflow of heat is described by an hysteresis loop whose segments AA' and BB', which correspond to a transition from one solution of the equations that are employed to another, are represented by arrows on the diagram.

Other segments of the hysteresis loop corresponding to stable glacial regimes can describe the dependence of

φ on $\frac{\Delta Q_{sp}}{Q_{sp}}$ for both increases and reductions in inflows of heat.

Let us note that the patterns represented in Figure 12 may also be established on the basis of general considerations that must be taken into account in constructing different models of the thermal regime.

It has already been noted that for the values of heat inflow to the atmosphere's outer boundary that exist today a stable existence of complete glaciation over our planet (a "white Earth") is possible, with very low temperatures at all latitudes.

The stability of such a regime is explained by the very large albedo of a surface covered by snow and ice. In such cases climatic conditions that are currently characteristic of the Antarctic region may develop over the entire surface of the Earth.

The conclusion that a stable existence of a "white Earth" is possible may apparently be inferred from any realistic theory of climate.

This means that for the existing inflow of solar radiation, a regime of full glaciation of the Earth, as indicated in Figure 12 by point 1, is possible. Such a regime is also possible for inflows of radiation that are less than the present ones (this is indicated on the diagram by the line moving to the left of point 1 along the horizontal axis), as well as if there is an increase in radiation inflow to values greater than those at present, up to the value that corresponds to the appearance of temperatures at which ice begins to melt in the warmest regions of the Earth. After such a temperature is reached a part of the Earth's surface may be freed from its ice cover. This will cause a reduction in the albedo and an increase in the absorbed radiation. The point that corresponds to the boundary of a regime of full glaciation under an increased solar radiation is represented by the letter *A*.

It is natural to assume that between the conditions represented by points *A* and 3 there exist regimes of partial glaciation of the Earth that may be represented on the diagram in the form of a line connecting these points. The shape of such a curve may be established in the following manner.

Both data from observations and physical considerations indicate that under current climatic conditions, as the inflow of heat to the Earth's surface increases the average air temperature at the Earth's surface rises and polar ice recedes. As the inflow of heat reduces one observes a reduction in the average temperature and an expansion of ice.

In accordance with that pattern the line connecting point *A* with point 3 should approach the latter point from the left, which is possible if this line intersects the vertical axis at least at one more point (point 2).

We thus conclude that it is possible for a third climatic regime to exist under contemporary conditions, which is a second variant of a partial glaciation of the Earth in which a larger area would be under ice cover than today. But that climatic regime cannot be viewed as stable, since it may pass to states of full glaciation or else of a full absence of ice cover following very small changes in values of climate-forming factors.

Following its intersection with the vertical axis at point 3, the line that describes regimes of partial glaciation must, for some value of incoming radiation exceeding its current magnitude, reach the horizontal line corresponding to an ice-free regime (point *E*). From point *E* the line that corresponds to a further increase in radiation becomes a horizontal straight line moving to the right.

In a number of the studies that have been mentioned, calculations were carried out that show that for currently existing fluxes of radiation an ice-free regime is possible in the Arctic region. If one accepts this conclusion, then one should represent such a regime on the diagram in the form of a point on the vertical axis for $\varphi = 90°$, and in such a case the line describing a regime of partial glaciation must reach the horizontal line corresponding to an ice-free regime not to the right of the end of the vertical axis, but to the left. Such a location for the point that links these lines is explained by the fact that if an ice-free regime is possible for existing inflows of radiation, then it is also possible for levels of radiation that are reduced by a very small amount.

Thus in the case under consideration the line that describes possible glacial regimes passes through the point that has been mentioned as a horizontal straight line, which means that that line will intersect the vertical axis at least at one more point. Accordingly. under conditions of a possible existence of an ice-free Arctic region for currently existing values of inflow of solar radiation, there must exist still another regime of partial glaciation with a smaller area of ice cover than under currently existing climatic conditions.

To supplement the relation presented in Figure 12 let us represent in Figure 13 the dependence of average global air temperature on the inflow of heat calculated with the help of the same model. (All notations in Figure 13 correspond to those in Figure 12.)

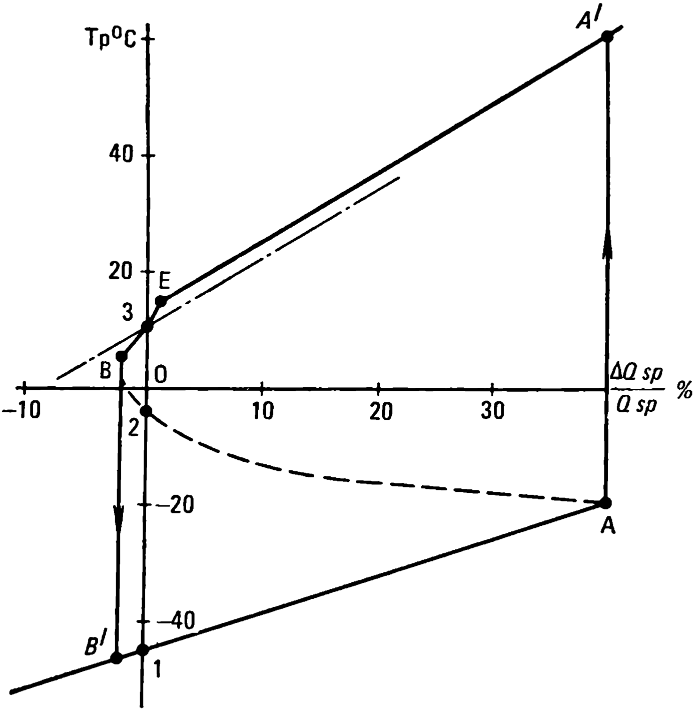

Fig. 13. Dependence of Average Temperature on the Inflow of Radiation to the Outer Boundary of the Atmosphere

It may be seen from Figure 13 that the dependence of the average global air temperature on the inflow of heat is in many ways similar to the corresponding dependence for the boundary of polar ice cover, and that it, too, may be described as an hysteresis loop. Actual changes in temperature in the course of the Earth's history appear to have been limited to short segments of the lines represented in Figures 12 and 13 which describe possible thermal regimes for our planet ranging from values occurring slightly above point *E* to a value lying somewhat above point *B*.

It is interesting to compare that part of the line under consideration in Figure 13 with the broken straight-and-dotted line passing through point 3, which corresponds to changes in the average global temperature in the absence of influences on the thermal regime from polar ices. Such a comparison shows the extent to which polar ices amplify changes in temperature produced by fluctuations in the inflow of heat.

It would thus appear that today's climate is not the only possible one for existing climate-forming factors. Aside from the existing climate, current external conditions may produce a climate corresponding to a "white Earth", as well as the other variants of climate mentioned earlier. This conclusion accords with the conception of ambiguity of climate developed by E. Lorenz (1968, etc.).

In order to estimate the stability of the current climate the relationships represented in Figures 12 and 13 may be employed.

It may be seen from these diagrams that the climatic regimes corresponding to segments *BE* are highly sensitive to very small changes in the inflow of heat. It follows that relatively small changes in external climate-forming factors may greatly alter the existing climate.

Following the publication of the first study devoted to the problem of the current climate stability (Budyko, 1968) that problem was considered in several tens of other studies in which both the model of the thermal regime of the atmosphere that has just been described and other similar semi-empirical models of climate theory were employed.

They have all confirmed this conclusion.

In such a context the study of the existing climate's stability carried out by R. Wetherald and S. Manabe (1975) is especially interesting.

Unlike the studies mentioned earlier, in which semi-empirical models of the distribution of average air temperatures among different latitudes were employed, Wetherald and Manabe applied a three-dimensional model of a general theory of climate that includes a detailed consideration of dynamic processes occurring in the atmosphere. That model takes into consideration the influence of state transformations of water on the thermal regime, including the feedback relation between the snow cover, polar ice, and air temperature.

Wetherald and Manabe have found that a reduction in the solar constant by more than four per cent extends snow and ice covers throughout the entire globe. An increase in the solar constant by two per cent causes a receding of the ice cover, as the average yearly air temperature at low latitudes increases by approximately 2°C. It increases by much more at high latitudes, and by as much as 10°C at 80° North latitude.

For comparison let us note that calculations based on the semi-empirical model of thermal regimes presented above show slightly larger changes in average yearly air temperature following increases in the solar constant by two per cent: at low latitudes it changes by 3-4°C and at 80° North latitude by 12-14°C.

It is also interesting to note that the findings of Wetherald and Manabe accord with the main conclusion obtained from the semi-empirical models of the thermal regime of the atmosphere concerning the great sensitivity of today's climate to small changes in inflows of heat arriving at the atmosphere's upper boundary.

The conclusions that were obtained in these studies revise earlier views concerning the unambiguity and high measure of stability of the contemporary climate.

3. Changes in Climate

Let us present a brief description of the history of changes in climate. This is based on paleogeographic studies for the geological past and on observations through networks of meteorological stations for the present time.

Climates in Prequaternary Periods. Little is known concerning climatic conditions in Precambrian times, and somewhat more information is available concerning climatic conditions of the Paleozoic era (570-235 million years B.C.). It would appear that during a major part of the Paleozoic era the climate was very warm in all parts of the globe, and that moisture conditions on the continents fluctuated within wide boundaries. Towards the end of the Paleozoic, at the boundary between the Carboniferous and Permian periods a glaciation developed that extended to a large part of the land areas that are today located at tropical latitudes. It is difficult to estimate the geographical location of that glaciation during its development since substantial shifts may have subsequently occurred in the location of the Earth's continents and poles.

Characteristically climatic conditions in other regions of the Earth during the Permocarboniferous glaciation were relatively warm.

During the Permian period a thermal zonality became perceptible, and on the continents regions of dry climate were greatly enlarged.

The climate of the Mesozoic era (235-66 million years B.C.) was relatively uniform. Over a large part of the Earth prevailing climatic conditions were similar to today's tropical climate, while at high latitudes the climate was cooler, though it remained quite warm, with negligible seasonal changes in temperature. Moisture conditions on the continents appear to have been homogeneous during the Mesozoic age by comparison with the present, even though zones of insufficient and of excessive humidity did exist.

Towards the end of the Cretaceous period the zone of hot climate became less extensive, while the region of dry climatic conditions spread. In passing to the Cenozoic no perceptible changes in climate occurred. During the second half of the Tertiary period (towards the middle of the Oligocene epoch) a process of progressive cooling began that was most pronounced at middle and especially at high latitudes. Since then a new climatic zone developed at high latitudes and gradually widened, whose meteorological regime was similar to that of current climatic conditions at middle latitudes. The winter air temperature in that zone fell below zero and this made possible the seasonal formation of snow covers. At the same time the specific properties of continental climates became more pronounced in continental regions distant from the ocean.

The cooling process was not uniform and there were periods of temperature rises as well. But these did not alter the general tendency towards an intensification of thermal zonality attributable to declining temperatures at high latitudes. That process was further intensified during the Pliocene epoch when the continental glaciation that started during the Oligocene in the Antarctic (and which continues to exist today) began to spread.

While towards the end of the Pliocene the climate became warmer than it is today, it was more similar to contemporary climatic conditions than to those of the Mesozoic era and of the first half of the Tertiary period.

Climatic Conditions in the Pleistocene Epoch. The climate of the Pleistocene epoch differed sharply from preceding conditions in the Mesozoic era and the Tertiary period when thermal zonality was not very pronounced. The Pleistocene epoch began 1.5-2 million years ago following an intensification of the cooling that occurred towards the end of the Tertiary period at middle and high latitudes. This contributed to the development of large continental glaciations. The number of such glaciations and their time of occurrence are known only approximately.

Studies carried out in the last century in the Alps make it possible to identify four principal European ice ages, which have been called Günz, Mindel, Riss, and Würm. Subsequently it became apparent that each of these glacial ages could be divided into several stages, and that glaciations receded during the intervals that divided them.

It appears that periods of development and of receding of glaciations occupied the smaller part of the Pleistocene, and that relatively warm intervening periods were more prolonged. At that time the ice cover on continents disappeared and continued to exist only in mountainous regions and at high latitudes. It has been established that the advance and receding of ice covers in Europe, Asia, and North America occurred more or less at the same time, and that a similar correspondence occurred in the ice ages in the Northern and Southern hemispheres.

The ice cover on continents spread furthest in regions possessing more humid maritime climates. In the relatively dry climates of Northern Asia glaciations occupied a relatively small area.

During periods of particularly vast glaciations the continental ice cover in the Northern Hemisphere reached, on the average, 57° North latitude, and in individual regions, 40° North latitude. The thickness of the continental ice cover over a major part of its area was several hundred meters, while in some regions it reached several kilometers.

Undoubtedly, as continental glaciations occurred the boundary of sea polar ice also moved to lower latitudes. This greatly increased the overall area of our globe's permanent ice cover. During each glaciation period the snow level in mountainous regions not subjected to glaciation moved down by hundreds of meters and sometimes more than a kilometer.

At the same time the zones of permafrost increased greatly during such ice ages. The corresponding boundary moved to lower latitudes over distances that sometimes reached several thousand kilometers.

A characteristic feature of ice ages was the development of extensive glaciations over the continents, and the reduction in the level of the World Ocean by 100-150 meters below its present level. During the warm periods between ice ages the level of oceans rose by several tens of meters over their current level.

The climate of ice ages was characterized by perceptible declines in air temperature in all regions of the world. This reduction in temperature which on the average amounted to several degrees below current values was more pronounced at higher latitudes. In warm periods between ice ages air temperature was higher than it is today, even though in this case the difference in air temperatures was less in absolute terms than during glaciation periods.

The influence of ice ages on precipitation is less clear. Some data suggest that moisture conditions in different parts of the world changed in different ways during periods of glaciation. This points to changes in the system of atmospheric circulation produced by glaciations and to corresponding changes in temperature differences between the Equator and the poles.

Climatic Conditions in the Holocene Epoch. The Holocene epoch represented a relatively short episode in the history of the climatic changes that followed the end of Quaternary glaciations. Nevertheless there were several perceptible fluctuations in climatic conditions during that epoch.

The maximal development of the last, Würmian, glaciation occurred approximately 20,000 years B.C., and several thousand years later this glaciation underwent substantial destruction. The epoch that followed was characterized by a relatively cold and humid climate at middle and high latitudes of the Northern Hemisphere. Approximately

12,000 years B.C. there occurred a substantial increase in temperature (the Allered) which was soon followed by a cool period. As a result of these climatic fluctuations Europe's summer air temperature changed by several degrees.

Subsequently temperature increased once again, and the last large-scale glaciations in Europe and North America disappeared 5-7 thousand years B.C. At that time post-glaciation temperature increases attained a maximum. It is assumed that between 5 and 6 thousand years B.C. air temperature at middle latitudes of the Northern Hemisphere was 1-3°C higher than at the present time.

At the same time changes evidently occurred in atmospheric circulation. While polar ice boundaries shifted to the north, the subtropical high-pressure belt moved to higher latitudes, and this led to an expansion of arid zones in a number of regions of Europe, Asia and North America. The volume of precipitation in contemporary desert areas at low latitudes also increased. During that period the Sahara's climate was relatively humid, and this made possible the existence of a varied flora and fauna. Subsequently a trend towards lower temperatures prevailed that was especially pronounced during the first half of the first millennium B.C. Together with changes in the thermal regime the precipitation regime changed as well and gradually approached its present state.

An appreciable rise in temperature occurred at the end of the first and the beginning of the second millennium A.D. At that time polar ice receded to high latitudes.

The process of cooling that began in the thirteenth century and reached a maximum at the beginning of the seventeenth century was accompanied by an extension of mountain glaciers, and is therefore sometimes called a small ice age. Subsequently the temperature increased again and the ice receded.

The climatic conditions of the eighteenth and nineteenth centuries differed little from those that exist today.

Contemporary Climatic Changes. In the age of instrumental observations the largest change in climate began towards the end of the nineteenth century. It was characterized by a gradual increase in air temperature at all latitudes of the Northern Hemisphere and in all seasons, but especially at high latitudes during cold seasons.

Increases in temperature reached a maximum in the 1930s when the average air temperature in the Northern Hemisphere increased by approximately 0.6°C by comparison with the end of the nineteenth century. In the 1940s this was replaced by a cooling process that continued until recently. That cooling was relatively slow and did not match the preceding cycle of increased temperatures.

Figure 14 represents the course of air temperature anomalies for the latitudinal zone between 70° and 85° North latitude and for the Northern Hemisphere.

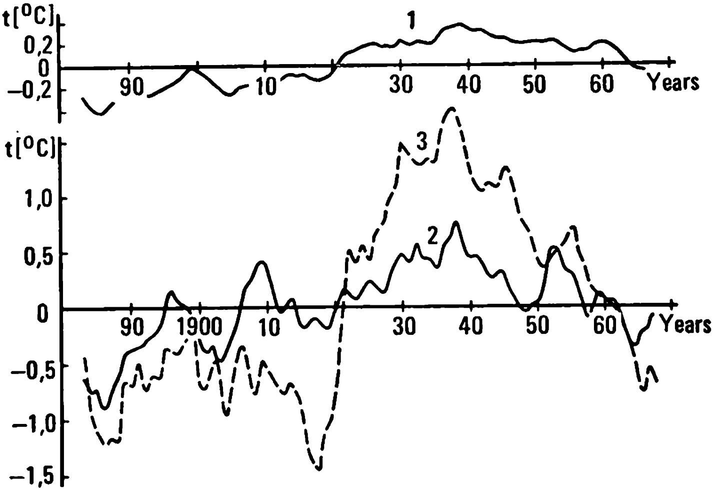

Fig. 14. The Secular Variations of Air Temperature Anomalies (Five-Year Running Averages):
1—anomalies in average yearly temperatures in the Northern Hemisphere; 2—anomalies in temperature in the latitudinal zone between 70° and 85° North latitude for warm seasons; 3—same for cold seasons

The unbroken line in the lower part of that diagram shows air temperature anomalies during the warm seasons while the broken line represents cold seasons. All data are averaged for the running five-year periods.

It follows from this figure that in the Northern Hemisphere a warming process began towards the end of the nineteenth century and attained an initial weakly pronounced maximum during the last years of that century. This was followed by a certain decline in temperature and then a rapid increase. The rate of increase became especially large during the cold seasons at the end of the 1910s and the beginning of the 1920s. The positive temperature anomaly attained a maximal value towards the end of the 1930s, while in 1940s this was replaced by a cooling process that became more intense in the 1960s.

In the middle of the 1960s the average temperature for extratropical latitudes of the Northern Hemisphere attained a level that corresponded approximately to the temperature at the end of the first decade. Thus the cooling process compensated for approximately one-half of the increase in temperature that occurred since the end of the nineteenth century.

While data concerning contemporary changes in climate in the Southern Hemisphere are less definite there are grounds for believing that during the first half of the twentieth century an increase in temperature occurred in the Southern Hemisphere as well.

In the Northern Hemisphere an increase of air temperature was accompanied by the reduction in the polar ice area, a receding of the boundary of permafrost to higher latitudes, and a northward movement of the forest and tundra boundaries as well as other changes in natural conditions.

An important role was played by changes in the regime of atmospheric precipitation that occurred in the period of increasing temperatures. The amount of precipitation in a number of regions of insufficient humidity was reduced as the climate became warmer, particularly during the cold seasons of the year. This may be seen, for instance, from Figure 15, which represents the total precipitation during this century for cold seasons (between November and March) in the steppe and forest-steppe zones in the USSR. The anomalies in precipitation were calculated for the running five-year periods.

The decline in precipitation has led to a decline in the river run-off and a reduction in the level of certain closed bodies of water. In particular the level of the Caspian Sea declined substantially in the 1930s largely as a result of the reduced run-off of the Volga River.

At the same time the frequency of droughts increased during the period of increased temperatures which extended to large areas in intracontinental regions at middle latitudes in Europe, Asia and North America. The resulting influence on the economies of a number of countries has been considerable.

Causes of Climatic Changes. Until recently studies of the causes underlying changes in climate were made difficult by the absence of a corresponding physical theory. In recent years, however, a major change has been brought about by numerical models of climate theory, including the semi-empirical theory of atmospheric thermal regime described earlier.

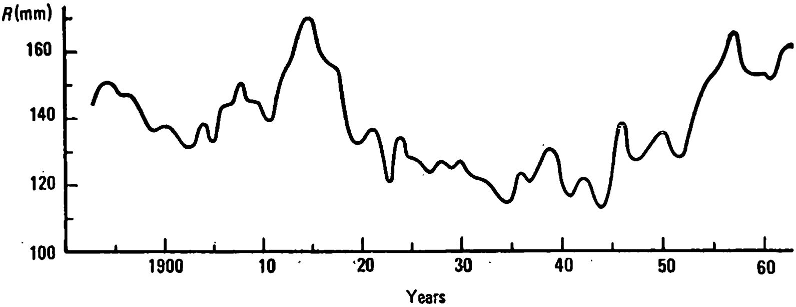

Fig. 15. Secular Variations of Total Precipitation for Cold Seasons in Steppe and Forest-Steppe Regions of the USSR

The sections that follow describe some of the findings resulting from such studies (Budyko, 1974).

Since quantitative data concerning climatic conditions in the distant past are not very reliable our survey of the causes of climate changes in Prequaternary times will be limited to an examination of the secular trend of air temperature during the second half of the Mesozoic era and the Tertiary period. We will rely on data collected by C. Emiliani (1966) and derived from a determination of paleotemperatures through methods of isotopic analysis.

The curve marked *1* in Figure 16 refers to the secular trend of the average temperature of the surface layer of oceans at middle latitudes determined by Emiliani. It shows that during the last 130 million years there has been a tendency for average air temperature at the Earth's surface to decline. That tendency was interrupted several times by perceptible increases in air temperature, but these were followed by periods of cooling and did not alter the overall course of the climate evolution.

In order to explain these changes in climatic conditions one should consider the influence on the thermal regime of fluctuations in the composition of the atmosphere and in the structure of the Earth's surface.

Let us consider the influence of the concentration of carbon dioxide within the atmosphere on the atmospheric thermal regime.

Data presented in Chapter X describe the prevailing tendency during the Cenozoic era towards a decline in the concentration of carbon dioxide in the atmosphere.

Since carbon dioxide influences the absorption of long-wave radiation, thus sustaining the atmospheric greenhouse effect, a reduction in its mass produces a decline in air temperature at the Earth's surface. It has been established that it is only for relatively low concentrations (less than 0.1 per cent) that the dependence of temperature

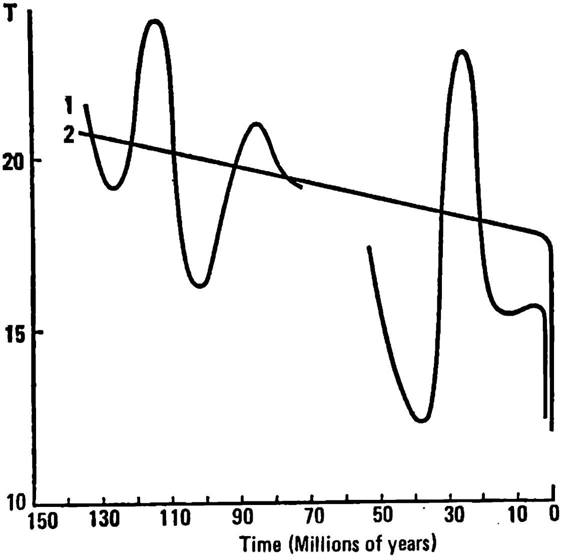

Fig. 16. Secular Variations of Average Temperature of Surface Layer of Ocean Waters at Middle Latitudes of Northern Hemisphere:
1—observed data;
2—calculated values

on the concentration of carbon dioxide becomes perceptible. At higher concentrations changes in the volume of carbon dioxide in the atmosphere influence air temperature insignificantly.

As a result it is only during the last several million years, that is during the Pliocene and Pleistocene epochs, that reductions in the content of carbon dioxide in the atmosphere have produced considerable changes in air temperature. Calculations based on the semi-empirical theory of a thermal regime described earlier have shown that during that period the average air temperature at the Earth's surface has declined by several degrees.

The atmospheric thermal regime has also been influenced by changes in the structure of the Earth's surface. During the second half of the Mesozoic era and during the first part of the Tertiary period (the Paleocene epoch) the average elevation of continents was relatively low, and as a result a large part of the continental platform was covered by shallow seas. During that time a large number of uplifts and sinkings of different continental areas occurred, but this did not alter the overall pattern in which straights and seas of various sizes separated various continents.

During the second half of the Tertiary period (the Neocene epoch) intensive tectonic movements produced the elevation of continents and caused a gradual vanishing of many intracontinental seas. In particular it was during that epoch that a large sea on the territory of present-day Western Siberia ceased to exist, that had linked tropical oceans with the polar basin. This has transformed today's Arctic Ocean into a relatively isolated body of water that is linked only with the Atlantic and poorly connected to the Pacific because of the narrowness of the Bering Strait.

It is possible that a slow drift of the Antarctic continent to higher latitudes was completed during the Tertiary period. Presumably such changes in the structure of the Earth's surface produced substantial declines in meridional heat exchanges in the oceans.

One can imagine that such reductions in meridional flows of heat in oceans resulting from increases in the levels of continents have contributed to the reduction in temperatures that have taken place at middle and high latitudes in the course of the last 100-150 million years.

The secular trend of temperatures at middle latitudes corresponding to line *2* in Figure 16 was calculated in such a way as to reflect the influence of changes in meridional heat exchanges in oceans on the thermal regime and the reduction of temperature that occurred towards the end of the Tertiary period as a result of reduced concentration of carbon dioxide. That line generally accords with the empirical curve marked *1*, even though the two curves do not fully coincide.

The reasons why the temperatures represented by curve *1* in Figure 16 fluctuate over intervals extending to tens of millions of years is an interesting problem. It may be assumed that this is due to local changes in circulation processes in the atmosphere and in oceans. Peculiar climatic conditions during the Quaternary period appear to have emerged because of the further decline of the carbon dioxide content in the atmosphere and because of continental drifts and their elevation.

The Quaternary period was preceded by a prolonged evolution in climate in the direction of a more pronounced thermal zonality resulting from changes in the Earth's surface structure. This was largely expressed in a continual decline in air temperature at middle and high latitudes.

During the Pliocene epoch climatic conditions began to be influenced by a reduction of the concentration of carbon dioxide in the atmosphere. This led to a decline in average global air temperature by 2-3°C (and 3-5°C at high latitudes). Subsequently polar ice caps extended, whose development led to a further reduction in the average global temperature, particularly at high latitudes.

The development of polar ice caps sharply increased the sensitivity of the thermal regime to even very small changes in climate-forming factors. This made possible very large oscillations in the boundaries of snow and ice covers on land and in oceans as a result of changes in the location of land in relation to the Sun; earlier this factor had not influenced the climate substantially.

A continuing reduction in the concentration of carbon dioxide in the atmosphere contributed to an advance of later glaciations, even though the major influence on their scope was a combination of astronomic factors determining the location of the Earth's surface in relation to the Sun.

These factors include the eccentricity of the Earth's orbit, the inclination of the Earth's axis in relation to the plane of its orbit and the time of equinoxes. These astronomic factors change periodically and produce certain changes in the quantity of radiation reaching different latitudes in various seasons. The periods within which these changes occur extend to tens of thousands of years. By comparison with changes in astronomic factors all other factors appear to have exerted a lesser influence on climatic fluctuations during the Quaternary period. In order to verify this conclusion one should make use of a numerical model that makes it possible to calculate the position of individual ice covers as influenced by external climate-forming factors.

To clarify this problem, use was made of the numerical model of the thermal regime for average yearly conditions prepared by the present author (Budyko, 1968). That model was described earlier. This has shown that changes in the radiation regime during the last, Würmian, glaciation could have led to a shift in the ice cover in the Northern Hemisphere to the south by a magnitude of approximately one degree latitude, which is substantially less than the shift in the ice cover that has in fact taken place.

In this connection we have noted that a determination of average yearly temperatures is insufficient to estimate the influence of changes in orbit elements on glaciation, since glaciation regimes are largely influenced by the thermal conditions of warm seasons.

In a subsequent investigation (Budyko, Vasishcheva, 1971) studies of climatic conditions during glaciation periods were based on a model describing the distribution of average latitudinal temperature in different seasons. Calculations were based on the method of successive approximations for the position of the average boundary of polar ice during periods when the influence of astronomic factors bearing on the radiation flux during warm seasons at high latitudes was substantially reduced. Some of the resulting findings are presented in Table 5. These show that fluctuations in the radiation regime caused by changes in the position of the Earth's surface in relation to the Sun may lead to substantial changes in climate. The corresponding calculations show that at such times average global temperature does not fluctuate much. Nevertheless such modest fluctuations are accompanied by perceptible shifts in the boundaries of ice covers.

It may be noted that the maximal changes in average latitudes reached by the ice cover within the Northern Hemisphere that are shown by these calculations accord well with empirical data. Thus, for example, the studies of H. Lamb (1964) have noted that at the height of glaciation the average boundary of ices in the Northern Hemisphere attained 57° North latitude, which in relation to current conditions represents a shift in that boundary by 15 degrees. The corresponding value established through calculations is 12 degrees, which is relatively close to the value noted by Lamb.

Table 5

CHANGES IN CLIMATE DURING GLACIATION PERIODS

Time (in thousands of years) until 1800 A. D.	$\Delta\varphi^{\circ}_{n}$	$\Delta\varphi^{\circ}_{s}$	ΔT°
22.1 (Würm III)	8	5	—5.2
71.9 (Würm II)	10	3	—5.9
116.1 (Würm I)	11	2	—6.5
187.5 (Riss II)	11	0	—6.4
232.4 (Riss I)	12	—4	—7.1

Note:

$\Delta\varphi^{\circ}_{n}$—reductions in the average latitude of the boundary of polar ice in the Northern Hemisphere by comparison with its current position

$\Delta\varphi^{\circ}_{s}$—reductions in the average latitude of the boundary of polar ice in the Southern Hemisphere by comparison with its current position

ΔT°—changes in the average temperature in warm seasons at 60° North latitude

One should consider the reasons explaining changes in climate that have occurred during the first half of our age. Studies of that problem have shown that the increased temperatures that attained a maximum value in the 1930s appear to have been caused by an increase in the transparency of the stratosphere, which has increased the inflow of solar radiation (the meteorological solar constant) entering

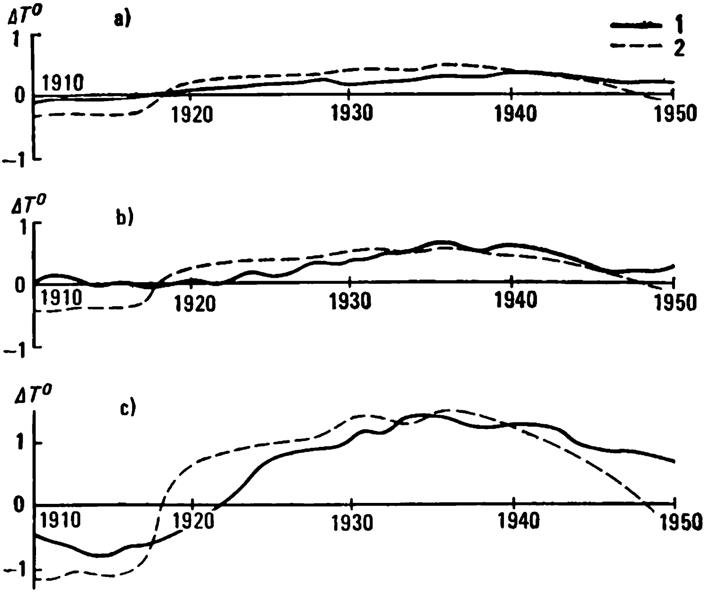

Fig. 17. Secular Variations of Air Temperature Anomalies: a—the Northern Hemisphere; b—zone corresponding to 70-80° North latitude during warm seasons; c—zone corresponding to 70-80° North latitude during cold seasons; 1—observed data; 2—calculated values (observed data are averaged for 10-year periods)

the troposphere. This has led to an increase in average global air temperature at the Earth's surface.

Changes in air temperature at different latitudes and in different seasons have depended on the optical thickness of the stratospheric aerosol and on shifts in the boundaries of sea polar ice. The receding of Arctic sea ice due to increased temperature has led to an additional perceptible increase in air temperature during cold seasons at high latitudes of the Northern Hemisphere.

These conclusions are confirmed by calculations based on the model of thermal regime of the atmosphere. These accord well with observed data as may be seen in Figure 17, in which the secular trend of air temperature anomalies in the Northern Hemisphere established through observations is compared with calculated values.

It seems probable that changes in the transparency of the stratosphere that have occurred during the first half of the twentieth century have been associated with the regime of volcanic activities, and in particular with changes in the inflow into the stratosphere of products of volcanic eruptions including sulphur dioxide.

There are grounds for believing that during the last 20 or 30 years changes in climate have begun to depend in some measure on human activities. This problem is considered in Chapter X.

IV

The Hydrological Cycle

1. The Water Balance

Water Balance Components. The biosphere's hydrological cycle includes exchange of water (in its various states) between the hydrosphere, atmosphere, upper layers of the lithosphere, and living organisms. The corresponding process may be described in terms of the equations of a water balance defined for the biosphere's individual components.

The water balance equation for land surfaces equates to zero the algebraic sum of all types of inflow and expenditure of water in solid, liquid and gas states entering a horizontal sector of the surface from surrounding space over a specified time interval.

That equation has the form:

$$r = E + f_w + G, \tag{4.1}$$

where r is precipitation, E—the difference between evaporation and condensation at the Earth's surface (this is usually simply called evaporation), f_w—surface run-off, and G—the flow of moisture from the Earth's surface to lower layers.

Equation (4.1) is generally employed in a somewhat modified form that may be obtained by taking into consideration the fact that the vertical flow of moisture G is equal to the sum of the run-off within the soil f_p and changes in the moisture content in upper layers of the lithosphere (b). That equality corresponds to the equation of the water balance for a vertical column passing through the upper layers of the lithosphere and reaching depths at which moisture exchange ceases to take place.

If we consider that the sum of surface run-off f_w and of run-off within the soil f_p is equal to the full run-off f we will have:

$$r = E + f + b. \tag{4.2}$$

Equation (4.2) may also be employed to calculate the water balances of bodies of water and of individual sectors within such bodies. In such a case f will describe the total redistribution of water along the horizontal plane over the period under consideration both within the body of water itself and in layers of the underlying soil (in those cases in which there does occur a significant redistribution of moisture at such levels). Similarly, for a closed water body b, too, will be equal to the overall change in the water content both within the body of water itself and in underlying layers in those cases in which there are perceptible changes in moisture content. In practice in many cases b describes changes in water levels. For an average yearly period b is often relatively small, and the water balance equation will then be:

$$r = E + f. \tag{4.3}$$

For the water balance of the Earth as a whole the horizontal redistribution of moisture is of no importance, and the water balance equation will have the simple form:

$$r = E. \tag{4.4}$$

This is also the form of the water balance equation for an average yearly period in the case of land surfaces which have no run-offs, such as deserts.

Finally let us consider the water balance equation for the atmosphere. By summing all categories of inflow and expenditure of moisture within a vertical column passing through the atmosphere it is not difficult to obtain the equation

$$E = r + C_a + b_a, \tag{4.5}$$

where C_a is a quantity of moisture that is received or else lost by a vertical column as a result of air currents and horizontal turbulent exchanges, and b_a—change in the quantity of water within that column.

Since the atmosphere can retain only relatively modest quantities of water in all its states, the value of b_a is usually much less than that of other components of the balance. Its average value over a year is always close to zero.

Water Balance of the Earth. Let us consider data concerning the components of the water balance of continents. The values of precipitation on land may be found from observations carried out by existing networks of meteorological stations.

The average yearly values of precipitation on different continents determined in the precipitation maps of L. P. Struzer and V. Ya. Sharova (*World's Water Balance*, 1974) are presented in Table 6. Table 6 does not include data on precipitation in the Antarctic and on all islands, though these are taken into consideration in calculating the average value of precipitation on all land.

Table 6

PRECIPITATION AND EVAPORATION ON CONTINENTS

Continent	Precipitation, cm/year	Evaporation, cm/year	Run-off, cm/year
Europe	77	49	28
Asia	63	37	26
Africa	72	58	14
North America	80	47	33
South America	160	94	66
Australia	45	41	4
All land	80	48.5	31.5

The instruments that are employed in measuring precipitation often record less than the actual value of precipitation, particularly in measuring the quantity of snow, because of wind snow does not get into rain gauge.

The resulting systematic errors in measuring precipitation are different for different measuring instruments. In constructing maps of precipitation attempts have been made to correct the more substantial errors of this type; yet the accepted adjustments were usually insufficient.

During the last several years it has been established that in regions with large amounts of solid precipitation the actual sums of precipitation may be larger than in earlier estimates by several tens per cent. The quantities of liquid forms of precipitation are usually measured with less systematic errors, though they, too, may occasionally be appreciable.

On the basis of studies of such errors in measuring precipitation L. P. Struzer and V. Ya. Sharova have constructed a world map of precipitation in which the adjustments have been taken into consideration to exclude the influence of major systematic errors in observing precipitation. For all land the corresponding data is nearly ten per cent higher than in earlier studies.

The data relating to run-off that are presented in Table 6 were obtained by generalizing available observations. This work was carried out in preparing *The World Water Balance.*

For each continent evaporation is equal to the difference between precipitation and run-off. The values of evaporation from the surface of continents were determined in the above study by L. I. Zubenok through calculations based on standard meteorological observations. They proved to be relatively close to the differences between precipitation and run-off.

It may be seen from Table 6 that the ratio of evaporation to run-off differs greatly on various continents. While in Australia evaporation is close to the magnitude of precipitation, in all other continents except Africa evaporation is less than two-thirds of the sum of precipitation.

By making use of L. P. Struzer's and V. Ya. Sharova's world map of precipitation, and of the world map of evaporation from oceans in *The World Water Balance* constructed by L. A. Strokina, the values of the water balance components for oceans may be calculated (Table 7).

It has already been noted that the difference between evaporation from the surface of the World Ocean and precipitation is equal to the magnitude of the river run-off from continents into the ocean. For individual oceans this is equal to the sum of the river run-off and the horizontal transfer of water from other oceans through circulation processes. It is difficult to determine the magnitude of that transfer through direct methods, since it represents a small difference between two values, namely the inflow and outflow of water, both subject to significant errors. It is somewhat simpler to estimate the exchange of water between oceans as a residual element in the water balance of each ocean, although in that case as well the accuracy of the corresponding estimates is limited.

Table 7

WATER BALANCE OF OCEANS

Ocean	Precipitation, cm/year	Evaporation, cm/year	Run-off, cm/year
Atlantic	101	136	23
Pacific	146	151	8
Indian	132	142	8
World Ocean	127	140	13

While Table 7 does not include data on the water balance of the Arctic Ocean (their accuracy is less than that of data for other oceans) these data are taken into consideration in estimating the components of the water balance for the World Ocean as a whole.

It may be seen from the table that the sum of precipitation and run-off for the Atlantic Ocean is less than the magnitude of evaporation. It follows that the Atlantic Ocean receives water from other oceans, including the Arctic Ocean, where evaporation is substantially smaller than the sum of precipitation and river run-off.

In the Indian Ocean the sum of precipitation and run-off is somewhat smaller than the magnitude of evaporation, while in the Pacific Ocean the sum of precipitation and run-off is larger than evaporation, which reflects a transfer of surplus water into other oceans.

Making use of these values for precipitation and evaporation for land and oceans, we find that for the Earth as a whole the magnitude of precipitation over a year, which is equal to that of evaporation, is 113 cm. Let us observe that this is more than has been estimated in all earlier studies.

By using world maps of precipitation and evaporation it is possible to determine the values of water balance components for various latitudinal zones. The results of such calculations show that the inflow of water vapour into the atmosphere from evaporation may be both larger and smaller in different latitudinal zones than its expenditure on precipitation. The source of water vapour in the atmosphere is primarily provided by high-pressure zones where evaporation greatly exceeds precipitation. That surplus of water vapour is expended in zones adjacent to the Equator and also at middle and high latitudes, where precipitation exceeds evaporation.

It is evident that the difference between precipitation and evaporation is also equal to the difference between the inflow of water vapour into the atmosphere and the outflow resulting from the horizontal air movement. The large values for that difference that are found in many regions are some indication of the importance of transfers of atmospheric water vapour in the formation of precipitation. The influence of transfers of water vapour on the volume of precipitation is examined in studies of hydrological cycle in the atmosphere, and some of the corresponding findings are presented in this chapter's concluding section.

The components of the continental and oceanic water balance are not constant: they change as a result of climatic fluctuations and other factors. Since changes in balance components over a year are small by comparison with their absolute values, they may be neglected in determining the corresponding magnitudes. In studies of the evolution of the hydrosphere, on the other hand, data on changes in the water balance components may be of considerable importance.

It was noted in Chapter I that the volume of water in the World Ocean changed perceptibly during glaciation periods, when large masses of water were expended on the formation of the continental ice sheets. Paleogeographic studies have established that during the age corresponding to the termination of the last glaciation, approximately twenty thousand years ago, the level of the World Ocean was lower than it is today by approximately one hundred meters. Subsequently that level gradually increased and has reached its current value about five thousand years ago. Since that time it has changed very little.

Systematic observations of the ocean level at a large number of locations were initiated in the late nineteenth century. They have shown that between 1900 and the early 1930s the World Ocean's level changed little. During the middle 1930s it increased at a rate of 0.2 cm per year. Altogether between 1900 and 1964 it has increased by approximately 10 cm.

It is not absolutely clear what sources of water have produced such an increase in the ocean's level. Possibly, during the period under consideration, a certain volume of water was lost by ice covers on land through melting processes (although the largest ice sheets, namely those in the Antarctic and in Greenland, appear not to have declined during that time). There are data indicating that in the twentieth century the reserves of subsoil water on land have declined, and also that the volume of water in many lakes has declined. Probably all the three factors have contributed to increasing the level of the World Ocean.

2. *Relations Between Energy and Water Balances on Land*

The run-off normal, that is the volume of water flowing off on the average during a year from a unit of land surface in the form of various horizontal flows, is a major characteristic of the hydrological regime of land. Another important indicator of hydrological conditions is the coefficient of run-off, i.e., the ratio of the run-off normal to total yearly precipitation.

Since the yearly run-off depends very much on the process of evaporation, which is one of the major processes in the transformation of solar energy at the Earth's surface, it is obvious that the run-off normal and run-off coefficient are linked to the principal components of the energy balance. An examination of that linkage should help clarify the principles that determine the specific features of the hydrological regime of different geographical zones.

The relation between elements of the energy and water balances of land was derived from the following considerations (Budyko, 1948). The average total evaporation from the surface of land E evidently depends on the quantity of precipitation r and the inflow of solar energy, and as precipitation and the radiation balance R increase so does evaporation. In conditions in which the soil is dry, the entire water arriving in the form of precipitation is caught by molecular forces on soil particles and is eventually expended on evaporation. In such cases (which occur, for example, in desert areas) the run-off coefficient $\frac{f}{r}$ approaches zero.

Since the average dryness of the soil increases when inflow of radiation heat increases and when the quantity of precipitation declines, it may be concluded that

$$\frac{f}{r} \to 0 \text{ or } \frac{E}{r} \to 1 \text{ when } \frac{R}{Lr} \to \infty. \tag{4.6}$$

When the value of the ratio $\frac{R}{Zr}$ is reduced (we will refer to it as the radiation index of dryness) the value of $\frac{E}{r}$ will decline (there will appear a certain run-off), and for values of total precipitation that are sufficiently large and total inflow of radiation that is sufficiently small a state of full moistening of the soil's upper layer will be achieved. In such cases the possible maximum of heat energy from available resources will be expended on evaporation. Its value may be estimated by considering the valve nature of turbulent heat exchanges between the underlying surface and the atmosphere.

Experimental studies have established that the turbulent heat conductivity of the lower air layer depends substantially on the direction of the vertical turbulent heat flow. When the turbulent flow moves away from the Earth towards the atmosphere a relatively larger intensity of turbulent mixing makes it possible for it to attain higher values which then approach those of the main components of the radiation and heat balances. For a reverse direction of the vertical turbulent heat flow inverting temperature distribution causes a decrease in the exchange intensity and as a result the turbulent heat flow is found to be relatively small. At middle latitudes in the winter, predominance of invertion conditions causes the turbulent heat exchange to be relatively small by comparison with its maximal summer values, which are determined by superadiabatic conditions of temperature distribution within the lower air layer. A similar regularity may be seen in daily cycles as well, by comparing the magnitudes of turbulent heat flows at night and in the daytime.

As a result of the valve effect yearly sums of turbulent heat flows are positive, that is the average turbulent heat flow moves from the Earth's surface to the atmosphere in nearly all climatic zones of land.

Concluding that yearly total turbulent heat flows cannot produce substantial inflows of energy to underlying surface we infer that expenditures of heat on evaporation are compensated only by the radiation balance, and that as a result the upper boundary for LE is equal to R. In other words it may be assumed that for conditions of full moistening

$$LE \to R \text{ when } \frac{R}{Lr} \to 0. \tag{4.7}$$

Conditions (4.6) and (4.7) determine the form of function Φ,

$$\frac{E}{r} = \Phi\left(\frac{R}{Lr}\right) \tag{4.8}$$

$$\text{for } \frac{R}{Lr} \to 0 \text{ and for } \frac{R}{Lr} \to \infty.$$

It should be noted that a relationship among the components of the water balance was established some time ago by P. Schreiber (1904) and E. M. Oldekop (1911) on the basis of data on precipitation and run-off. This was described by Oldekop through the formulae:

$$E = r\left(1 - e^{-\frac{E_0}{r}}\right) \tag{4.9}$$

(the Schreiber-Oldekop equation, where E_0 represents the maximum possible value of evaporation under given conditions) and

$$E = E_0 \, th \frac{r}{E_0} \tag{4.10}$$

(the Oldekop equation, where *th* is a function for the hyperbolic tangent).

It may be easily shown that both these formulae will meet conditions (4.6) and (4.7) if, in accordance with the considerations mentioned earlier, it is assumed that $E_0 = \frac{R}{L}$.

It should be kept in mind that the magnitude of possible evaporation at a given locality will be determined by the radiation balance corresponding to conditions of sufficient moisture at which vegetation can exist. The value of the radiation balance for conditions of sufficient moisture may differ from its value at the same locality when moisture is insufficient.

Let us note the reasons why the radiation balance of an active surface depends on moisture conditions. In most geographic regions possessing a more or less humid climate the albedo of the Earth's surface changes relatively little as a result of changes in humidity. Since in the presence of sufficient moisture the average difference in temperature between the Earth's surface and the air is relatively small, the approximate magnitude of evaporation may be inferred from the radiation balance corresponding to the actual state of the Earth's surface.

A different situation exists in dry climatic conditions when following the increase in humidity in a given locality changes occur in the albedo and the Earth's surface temperature, which approaches the air temperature in conditions of sufficient moisture. It is evident that in order to determine evaporation one should then make use of the value for the radiation balance corresponding to the albedo of a moist surface and to a temperature at the Earth's surface that is equal to air temperature.

In order to verify these considerations concerning the nature of the relation between $\frac{E}{r}$ and $\frac{R}{Lr}$ for both small and large values of the last parameter different empirical data were employed in the studies mentioned earlier. As an example let us cite a comparison of the values of the radiation index of dryness $\frac{R}{Lr}$ with the values of the ratio $\frac{E}{r}$, for data corresponding to river basins on different continents (excluding mountainous regions).

On Figure 18 the ratio $\frac{E}{r}$ for low values of $\frac{R}{Lr}$ is represented in accordance with condition (4.7) as a straight line *OA*, while for large values of $\frac{R}{Lr}$ it is represented in accordance with condition (4.6) as a straight line *AB*. The points representing values of the ratio $\frac{E}{r}$ are calculated from water-balance data by averaging $\frac{E}{r}$ for specific intervals of the parameter $\frac{R}{Lr}$.

The experimental points show that in fact there exists a smooth transition from line OA to line AB which are, as expected, limiting values for the relation between $\frac{E}{r}$ and $\frac{R}{Lr}$.

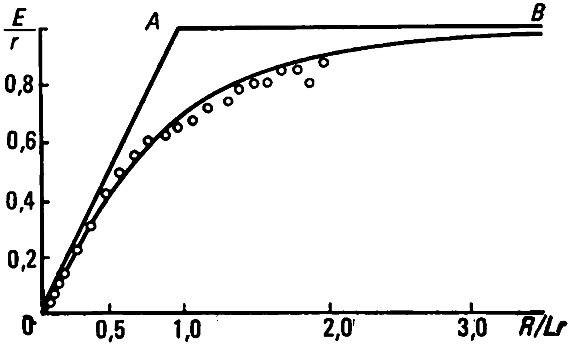

Fig. 18. The Dependence of the Evaporation-Precipitation Ratio on the Radiation Index of Dryness

In order to represent the relation (4.8) analytically, formulae similar to (4.9) and (4.10) may be employed:

$$E = r\left(1 - e^{-\frac{R}{Lr}}\right) \tag{4.11}$$

and

$$E = \frac{R}{L}\, th \frac{Lr}{R}. \tag{4.12}$$

One may also employ the geometric average of the right-hand terms of these equations:

$$E = \sqrt{\frac{Rr}{L}\, th \frac{Lr}{R}\left(1 - ch \frac{R}{Lr} + sh \frac{R}{Lr}\right)}, \tag{4.13}$$

where *ch* and *sh* are hyperbolic cosine and sine. In Figure 18 this is represented by a curve that fits most of the experimental data somewhat better than do the first two relationships.

Equation (4.13) represents the relation between the average yearly evaporation and precipitation, and the radiation balance.

Since run-off normal is equal to the difference between precipitation and evaporation, it follows that this formula also makes it possible to calculate run-off from data on precipitation and the radiation balance.

Equations representing the run-off and run-off coefficient in terms of formula (4.13) will then be:

$$f = r - \sqrt{\frac{Rr}{L}\, th\, \frac{Lr}{R}\left(1 - ch\, \frac{R}{Lr} + sh\, \frac{R}{Lr}\right)} \qquad (4.14)$$

and

$$\frac{f}{r} = 1 - \sqrt{\frac{R}{Lr}\, th\, \frac{Lr}{R}\left(1 - ch\, \frac{R}{Lr} + sh\, \frac{R}{Lr}\right)}\,. \qquad (4.15)$$

Let us consider the physical content of the equation linking the energy and water balances. As may be seen in Figure 18 the relationship on which that equation is based is largely determined by two limiting conditions, one of which is based on a representation of the valve mechanism of turbulent heat exchange in the layer of air adjoining the Earth, while the second corresponds to the obviously small value of the coefficient of run-off in dry climates. The selection of one or the other interpolation function for passing from the first of these conditions to the second is not particularly significant, since for most of the interval of changes in the parameters of the linkage equation the corresponding relation remains close to a boundary condition.

Thus even though the equation represents a semi-empirical relationship the role of empirical data in its justification is small by comparison with the general considerations that were mentioned earlier. It should be noted that the linkage equation represents supplementary relation that is independent of the energy balance and water balance equations.

Formulae (4.14) and (4.15), which establish a relation between components of the energy and water balances, have been verified in a number of studies. Of these we will mention the calculation of coefficients of run-off for the rivers of Europe (Budyko, 1951). For 29 rivers with basins whose area was more than 10,000 sq km, the average error in determining the coefficient of run-off was 0.04, and the values of that coefficient ranged from 0.13 to 0.64. The calculation error of run-off coefficient for rivers whose run-off coefficients are relatively large (more than 0.30) was 7 per cent. This is close to the accuracy of measurement of both precipitation and run-off. It follows that for large basins whose coefficients of run-off are not too small and for existing level of accuracy in determining precipitation and run-off it is sometimes impossible to establish the difference between "climatic" and real run-off.

A more comprehensive verification of the equation linking energy and water balances on land was carried out on the basis of data for 1,200 land regions, for which evaporation was determined in terms of the water balance

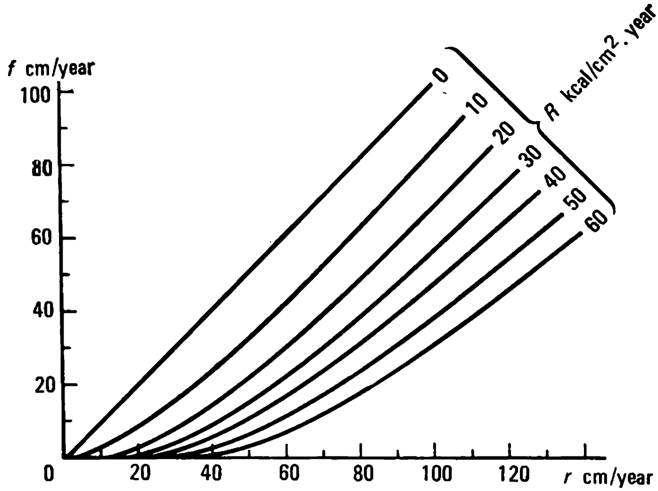

Fig. 19. The Dependence of Run-off on Precipitation and the Radiation Balance

as a difference between precipitation and run-off (Budyko, Zubenok, 1961). It was found that the average discrepancy between ratios of evaporation to precipitation calculated in terms of the linkage equation and those established in terms of the water balance was 10 per cent.

It was also noted in that study that in certain cases there are systematic deviations from calculated average linkage equation values, whose nature depends on the seasonal course of evaporation and precipitation. In cases in which changes in monthly values of evaporation and precipitation were parallel, average yearly values of the ratio of evaporation to precipitation increase somewhat, while when increases in evaporation in the seasonal course are matched by reductions in precipitation, average values of the ratio of evaporation to precipitation decline. In most cases these deviations lie within the range of accuracy of the values of evaporation as calculated from the linkage equation.

The linkage equation makes it possible to represent the relation of run-off and evaporation to total yearly precipitation and the radiation balance in a general form. The corresponding graph is shown in Figure 19. This regularity explains a number of empirical relations between run-off and precipitation that were found in earlier studies.

The linkage equation makes it possible to express the dependence of run-off on precipitation for values of the radiation balance corresponding to particular localities. This is shown by curve A in Figure 20 which was calculated for average conditions in the European lowlands. For purposes of comparison Figure 20 also shows the empirical relation found by H. Keller (1906), which relates run-off normal to precipitation on the basis of observation at West European rivers (line B), and the empirical curve by D. L. Sokolovsky (1936), which relates run-off to precipitation on the basis of observations on East European rivers (curve C). The correspondence of the linkage equation to these

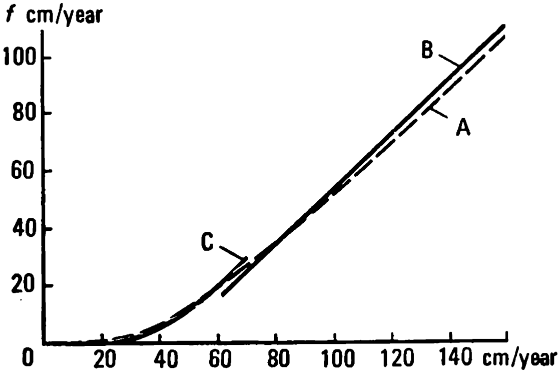

Fig. 20. The Dependence of Run-off on Precipitation (the horizontal axis represents the volume of precipitation r)

relationships represents still another confirmation of its universal character.

The linkage equation also makes it possible to explain the considerable dispersion that exists in dependence of run-off on precipitation in data from various regions which became evident in calculating average empirical relations (Polyakov, 1946). The substantial variability of the radiation balance at middle latitudes causes the run-off of basins with large radiation balances (i.e., in the more southerly regions) to be much smaller than that of basins with a low radiation balance (i.e., in the more northerly regions), for equal total values of precipitation. This also influences the rate of change of run-off corresponding to increases in precipitation $\frac{df}{dr}$. According to the linkage equation, it should be larger in northern basins than in southern ones. This is amply confirmed by empirical data.

The correspondence of run-off calculations based on the equation linking energy and water balances with observed data confirms the decisive role of climatic factors (and in particular of energy factors) in total yearly run-off. It is important to emphasize that this conclusion relates to river basins with areas large enough to be compared with the geographical zones. In the case of small areas, river run-off may change substantially as a result of local conditions of a nonclimatic nature.

3. The Water Exchange in the Atmosphere

Water Exchange Theory. Among the various forms in which living organisms influence the climate at any given time the transpiration of plant covers is especially important. Since transpiration usually represents several tens per cent of total evaporation, changes in transpiration may exert a considerable influence on the atmospheric water exchange, and hence on the volume of precipitation.

In studying this influence it is important to clarify the principles that govern the atmospheric water exchange.

The first studies of the water exchange were carried out at the beginning of this century. Their authors assumed that in the case of land, in each region the volume of water vapour brought from the outside, which is expended on the precipitation formation, does not exceed the difference between precipitation and evaporation.

Since in fact the overall volume of precipitation on land is generally much larger, these authors concluded that there exists a repeated internal water exchange. They assumed that the water vapour brought from outside into a particular land region precipitated within that region several times before being carried further by atmospheric circulation. It follows from such a conception of repeated internal water exchange that substantial changes in total precipitation could result from relatively modest changes in evaporation from the surface of land.

The idea of repeated water exchange was disproven as a result of studies that produced the quantitative theory of water exchange (Budyko, Drozdov, 1950; Kashin, Pogosian, 1950, and others). We will describe the principal equations representing the water exchange (Budyko, Drozdov, 1953).

Let us consider the transfer of water vapour in the atmosphere over a specified territory whose average linear scale is equal to L. The flow of water vapour brought by air currents to the given territory may be assumed to equal wu, where w represents the atmospheric moisture content in the windward side of the area under consideration and u is the average velocity of air flows carrying water vapour over the territory. The atmospheric moisture content will change along the air current's path in accordance with the difference between water expenditure on precipitation and inflow from evaporation.

It is evident that the flow of water vapour carried by air currents from the given territory will be equal to $wu-(r-E)L$, where r is total precipitation, E—total evaporation over the specified period of time, and L—the territory's linear scale.

The overall flow of water vapour that is carried over a given territory is made up of two subflows, namely one of external (advective) water vapour, which is formed by evaporation processes outside the given territory, and a local subflow formed by local evaporation processes.

The first flow on the area's windward side will be equal to wu, while on the leeward side (as the flow leaves the area's boundaries) it will be equal to $wu-r_aL$, where

r_a is total precipitation produced from external (advective) water vapour. The second flow is equal to zero on the windward side, while on the leeward side (as the flow leaves the given territory) it is equal to $(E-r_l)\,L$, where r_l represents total precipitation produced from water vapour of local origin.

Thus on the average for a given territory there are two flows, namely a flow of external water vapour equal to $wu-{}^1/_2 r_a L$, and a flow of local water vapour ${}^1/_2\,(E-r_l)L$. Together they produce an overall flow of $wu-{}^1/_2\,(r-E)\,L$. In this connection one should keep in mind that $r_a+r_l=r$. Since the molecules of the water vapour of local origin and external origin are mixed in the atmosphere in the course of turbulent exchange, it is evident that the ratio of total precipitation deriving from local and external water vapour is equal to the ratio of the quantity of the corresponding molecules of water vapour in the atmosphere.

In short it may be assumed that

$$\frac{r_a}{r_l}=\frac{wu-\frac{1}{2}r_a L}{\frac{1}{2}(E-r_l)\,L}, \qquad (4.16)$$

This yields the following two equations:

$$r_a=r\frac{1}{1+\frac{EL}{2wu}} \qquad (4.17)$$

and

$$r_l=r\frac{1}{1+\frac{2wu}{EL}}. \qquad (4.18)$$

From relation (4.17) it is also possible to determine the magnitude of the coefficient of water exchange K, which is equal to the ratio of the total sum of precipitation to the volume of precipitation of external (advective) origin:

$$K=\frac{r}{r_a}=1+\frac{EL}{2wu}. \qquad (4.19)$$

With the help of these formulae it is possible to analyze the dependence of the characteristics of water exchange on the basic factors that influence it. In particular, it follows from formula (4.19) that the coefficient of water exchange depends on the factors determining the water vapour balance in the atmosphere and does not depend directly on the magnitude of the river run-off that was utilized in certain past calculations of the coefficient of water exchange. Let us also note that the relation expressed in formulae (4.17), (4.18) and (4.19) between external precipitation, local precipitation and the coefficient of water exchange depends on the size of the territory under consideration.

Table 8

SEASONAL COURSE OF WATER EXCHANGE COMPONENTS FOR THE EUROPEAN PART OF THE USSR

	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII	Year
E, mm/month . . .	5	5	10	36	50	54	50	39	22	11	7	5	294
w, cm	0.4	0.4	0.6	0.9	1.5	2.0	2.3	2.2	1.6	1.2	0.8	0.5	1.2
u, m/sec	7.7	7.8	7.8	7.2	6.6	6.2	5.8	6.3	6.9	7.5	7.7	7.6	7.1
K	1.07	1.08	1.09	1.24	1.22	1.19	1.17	1.12	1.08	1.05	1.05	1.06	1.12
r, cm/month	2.7	2.3	2.4	2.8	3.8	5.5	6.3	5.9	5.1	4.9	3.8	3.2	48.7
r_a »	2.5	2.1	2.2	2.3	3.1	4.6	5.4	5.3	4.7	4.7	3.6	3.0	43.5
r_l »	0.2	0.2	0.2	0.5	0.7	0.9	0.9	0.6	0.4	0.2	0.2	0.2	5.2

As its scale L increases, so do total local precipitation and the coefficient of water exchange, while the total precipitation attributable to water vapour of external origin declines. Moreover, for territories that are not too small the actual dependence of the coefficient of water exchange on their size is not linear: for larger territories the average rate of transfer of water vapour u declines somewhat as a result of the curvature of air particle trajectories.

In estimating the influence of transfer of water vapour in the atmosphere on precipitation let us consider calculated data for the water exchange components in the European part of the USSR (see Table 8). The table shows that precipitation derived from local water vapour represents a very small part of the overall volume of precipitation. Both for the year as a whole and for individual months within the year the coefficient of water exchange is only slightly greater than a unit. This contradicts the view that there exists a multiple internal water exchange.

Actually even on such a large territory as the European part of the USSR only a very small part of the total volume of precipitation (approximately 12 per cent) falls a second time using local water vapour sources. The major part originates in water vapour brought from the outside.

O. A. Drozdov and his associates have studied the atmospheric water exchange over all continents (*World Water Balance*). Their findings are presented in Table 9. These show that even in the case of the largest continent, namely Asia, the larger part of precipitation is formed from water vapour brought to the continent from the outside. Thus the principal source of moisture for various continents is oceanic water vapour.

Table 9

ATMOSPHERIC WATER EXCHANGE OVER CONTINENTS

Continent	Precipitation, km^2/year	Advective precipitation, km^2/year	Local precipitation, km^2/year	Coefficient of water exchange
Europe	7,540	5,310	2,230	1.42
Asia	33,240	18,360	14,880	1.81
Africa	21,400	15,080	6,330	1.42
North America	16,200	9,790	6,360	1.65
South America	28,400	16,900	11,500	1.68
Australia	3,470	3,040	430	1.14

Because of the large scale of transfer of oceanic moisture in the atmosphere the contribution of local evaporation processes from the surface of continents to the total volume of precipitation is relatively modest, especially for territories whose dimensions are less than 1 million km^2. Yet this does not imply that the influence of local evaporation on total precipitation is limited to changes in the components of atmospheric water exchange: in addition to the direct influence of local evaporation on water exchange there is an indirect influence as well, which derives from the linkage between the volume of precipitation and the relative air humidity.

The studies of O. A. Drozdov (Drozdov, Grigoriyeva, 1963) have produced a semi-empirical formula for establishing the volume of precipitation:

$$r = aw\varphi(h), \tag{4.20}$$

where h is the average relative air humidity within the layer 0-7 km, w—the atmospheric moisture content, and a—a coefficient of proportionality that is equal to 1 when for h equal to 100 per cent a volume of precipitation of 0.1 w falls per day. That coefficient is close to 1 for the European part of the USSR.

The empirical function $\varphi(h)$ is generally similar for many regions at middle latitudes. In particular for the European part of the USSR the function $\varphi(h)$ is described by the data in Table 10.

Table 10

THE FUNCTION $\varphi(h)$

h, %	40	45	50	55	60	65	70	75	80	85	90	95	100
$\varphi(h)$	0.7	1.4	3.7	5.7	6.8	7.5	8.2	8.6	9.1	9.5	9.7	9.9	10.0

It may be seen from this table that precipitation is negligible for relative humidity below 40 per cent and that its magnitude increases rapidly as the relative air humidity increases to 50-55 per cent; for subsequent increases in humidity the increase in precipitation is less rapid, and for conditions of high air humidity its changes do not influence precipitation substantially.

Formula (4.20) may be employed to establish the dependence of precipitation on local evaporation processes. The values of w and h are then determined in accordance with the influence of local evaporation on the overall quantity of water vapour that is being carried over the given territory. In this connection it becomes apparent that the influence of local evaporation on precipitation within specific limits of air humidity may be several times larger than the contribution of local evaporation processes to the overall volume of atmospheric water vapour.

Thus the theory of atmospheric water exchange which describes the origins of water falling in the form of precipitation provides a basis for considering only the direct influence of local evaporation on the volume of precipitation. But as the studies of O. A. Drozdov have shown, a large role is often played by the indirect influence of local evaporation on precipitation, without which continental regions distant from the ocean would have been transformed into deserts.

Formula (4.20) makes it possible to calculate the dependence of volume of precipitation on the plant cover, which influences the quantity of local evaporation. But the relation between total evaporation and the character of vegetation has not yet been studied sufficiently.

In particular, even though the influence of forests on evaporation has been discussed for many years its nature is still not clear. Very probably forests do not influence total evaporation very substantially by comparison with other types of plant cover. But when mossy swamps are drained and a forest cover develops in such areas, total evaporation does appear to increase, and the reverse effect may take place when lowland grassy swamps are drained.

It may be assumed that the replacement of various types of plant cover changes total evaporation by approximately 10 per cent. With the help of formula (4.20) we find that the influence of changes in evaporation on precipitation will depend on the size of the territory within which evaporation has changed. If this has occurred within a geographical zone whose linear scale is of the order of 1,000 km, the influence of fluctuations in evaporation on precipitation will be negligible. But this influence may become substantial in the case of evaporation changes over entire continents or large parts of continents.

It is likely that during the Lower Paleozoic era, when there was no plant cover on the continents, the ability of upper layers of the lithosphere to retain water from precipitation was relatively small. Under these conditions evaporation from land was reduced and run-off increased. Because this proportion among components of the water balance existed over large areas it exerted a substantial influence on the volume of precipitation over continents, The appearance of plants on land surfaces and the development of soils increased evaporation and reduced run-off. As the land area under plant cover increased so did the volume of atmospheric precipitation, and that in turn contributed to the further spread of plants within continents. Such a process is an example of a positive feedback relationship between organisms and their environment: it contributed to a spreading of the plant cover over a major part of the surface of continents.

At the same time this particular feedback relation could alter precipitation perceptibly only in regions possessing favourable conditions of atmospheric circulation. The desert regions that exist today in subtropical high-pressure zones and elsewhere, as well as deserts that have existed in the past illustrate the limited influence of plant covers on global climate.

V

Geographical Zones

1. Climatic Factors Determining Geographical Zonality

Geographical Zonality. It was Dokuchayev who first drew attention to the existence of geographical zones, noting that within wide territories (zones) natural conditions retain many common features that change perceptibly from zone to zone. He noted that under the influence of the Earth's spherical shape climate, plants, and animals are distributed from north to south over the Earth's surface in a strictly determined order. And since the distribution of soil-forming agents is governed by specific principles and is related to specific belts, their product, namely soils, must also be distributed along the Earth's surface in terms of specific zones that follow latitudinal circles with only small deviations (Dokuchayev, 1948).

Dokuchayev identified the following major natural zones, and provided each with a short description: the Boreal or Tundra zone, the North Forest zone, the Forest-Steppe zone, the Steppe zone, the zone of Arid Steppes, the Desert zone, and the Laterite zone. (The names given to these zones vary somewhat in his various studies.)

Among subsequent studies of geographical zones those of A. Grigoriyev deserve particular attention. They include a series of studies, subsequently published as a collection (Grigoriyev, 1970), in which he considers the spatial structure of the geographical environment.

Grigoriyev identified four types of geographical environments that are associated with different latitudinal belts, Of these the first is the Equatorial belt, in which high air temperatures and abundant precipitation are found throughout the year. That belt's characteristics include the development of humid tropical forests, high run-off coefficient, and intensive erosion processes.

The second belt is the Tropical belt, which is located at latitudes at which the Sun's height differs in summer and winter months, thus producing a pronounced yearly air temperature cycle. In that belt, too, the Earth-atmosphere system's radiation balance is positive. The volume of precipitation in the Tropical belt fluctuates within wide boundaries, and this contributes to the development of three different zones on its territory. Within the external tropical zone total precipitation is small, and as a result one finds there semi-desert and desert areas with sparse drought-resisting plants, a negligible run-off coefficient, and intensive physical weathering. In the zone adjoining the Equatorial belt natural conditions are similar to those of the Equatorial belt in the summer but in the winter they are similar to those of the external tropical zone. This produces sharp changes in moisture conditions during the year, and in a number of regions a second less pronounced dry period occurs in the middle of the summer season. The plant cover of the zone adjoining the Equatorial belt is diverse and includes light tropical and savanna forests (deciduous and coniferous), as well as open savannas. The river run-off in that zone changes during the year in accordance with changes in precipitation, and the weathering process is seasonal in character. In particular, chemical erosion processes prevail during rainy seasons while physical erosion prevails during dry seasons.

The third zone of the Tropical belt identified by Grigoriyev as intermediate is characterized by a prevalence of drought conditions during the year with a relatively short humid period, when a limited volume of precipitation occurs. In a number of respects that zone is similar to the external tropical zone, although conditions are far more favourable to the development of plant life.

The next latitudinal belt according to Grigoriyev includes temperate and subtropical latitudes, where yearly magnitudes of the Earth-atmosphere system's radiation balance are negative but air temperature in the warmest months does not fall below 5°C. Because of differences in conditions of atmospheric circulation the volume of precipitation within that belt ranges from very high values to values that are close to zero.

The various combinations of heat and moisture regimes in the third latitudinal belt correspond to differences in the types of plant cover—tundras, forest-tundras, coniferous, mixed and deciduous forests, forest-steppe, steppes, semi-deserts, and deserts. Depending on moisture conditions the magnitude of run-off coefficients ranges from zero to values that are close to a unit. The process of weathering changes substantially during the year under the influence of the yearly cycle of temperature.

The fourth and last, latitudinal land belt in the Northern Hemisphere has been named Arctic by Grigoriyev. It is located at high latitudes, where low temperatures prevail throughout the year. Within that zone sparse plant life is found only in peripheral regions that enclose areas of permanent snow and ice—a desert of ice. In regions that are free of ice excessive moisture prevails together with a high run-off coefficient.

In a number of studies, subsequently published as a collection (1966) Grigoriyev has examined the patterns of the formation of geographical zones. The findings of some of his studies which were carried out jointly with the present author are presented below.

Climatic Indices Characterizing Geographical Zones. Dokuchayev observed that the boundaries of geographical zones are largely determined by climatic factors, and that in particular they depend substantially on moisture conditions. His studies contained estimates of ratios of precipitation to potential evaporation for major geographical zones. This marked the beginning of numerous studies that were devoted to an examination of the distribution of soils and plants depending on the ratios of precipitation to potential evaporation.

Among such studies one should mention that of G. N. Vysotsky (1905), which contains calculations for a number of natural zones of the ratio of yearly precipitation to potential evaporation (which refers to rates of evaporation from water surfaces). Vysotsky then made use of these values in analyzing conditions bearing on the formation of different types of soil and plant cover in various natural zones.

A different approach to the study of moisture conditions based on the utilization of data relating to the radiation regime was proposed by Grigoriyev (1946, 1954, etc.).

The following observation was made in our own study (Budyko, 1948): the equations for the mean yearly heat and water balances of land may be written in the form

$$\frac{R}{Lr}=\frac{E}{r}+\frac{P}{Lr}, \tag{5.1}$$

$$1=\frac{E}{r}+\frac{f}{r} \tag{5.2}$$

(the members of the heat balance equation are divided by Lr, and of the water balance by r.)

The equation linking the water and heat balance may then be added to these relations:

$$\frac{E}{r}=\Phi\left(\frac{R}{Lr}\right). \tag{5.3}$$

These three equations link four relative values of components of the heat and water balances. It is therefore sufficient to know any one of them in order to determine the others.

In view of the special form of the linkage equation, it follows that the ratio $\frac{R}{Lr}$ or else $\frac{P}{Lr}$ may be selected as the parameter that determines all relative values of components of the heat and water balances $\left(\text{the ratio } \frac{E}{r}\text{, or else } \frac{f}{r}\text{, is not a decisive factor for the first two variables in the case of dry climates, when small changes in the ratio } \frac{E}{r} \text{ or } \frac{f}{r} \text{ produce large changes in } \frac{R}{Lr} \text{ or } \frac{P}{Lr}\right)$.

Estimates of the magnitude of the turbulent flow are usually less accurate than the radiation balance, and it would seem more appropriate to select $\frac{R}{Lr}$ as the principal parameter that determines the relative values of components of the heat and water balances. That parameter, moreover, may be viewed as the ratio of potential evaporation $\frac{R}{L}$ to precipitation r, or else as a ratio of the radiation balance of a moistened surface to heat expended on the evaporation of the total yearly precipitation.

Such a use of that parameter thus embodies both Dokuchayev's and Vysotsky's idea of an indicator of moistening and Grigoriyev's view concerning the relevance of the ratio of the radiation balance and precipitation to the characterizing of moisture conditions.

While the relative values of components of the heat and water balances are determined by one parameter $\frac{R}{Lr}$, the absolute values of these components require two parameters, namely $\frac{R}{Lr}$ and R.

In order to clarify the physical meaning of the radiation index of dryness one should consider the relation between potential evaporation and the radiation balance of the Earth's surface.

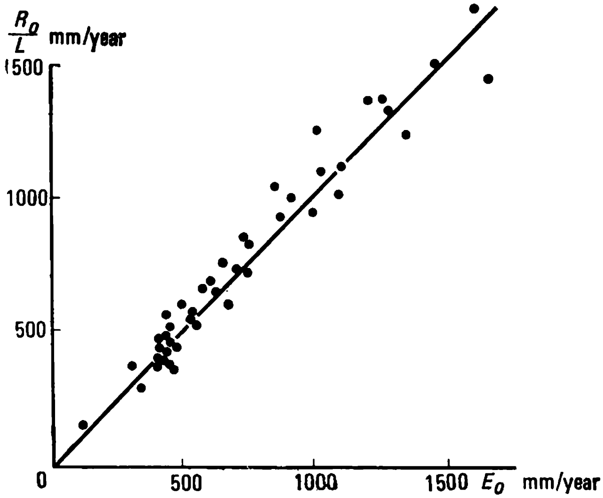

Fig. 21. A Comparison of Potential Evaporation Values Determined Through the Complex Method and from the Radiation Balance

The magnitude of potential evaporation from the surface of land under conditions of sufficient moisture may be determined in accordance with the formula describing evaporation from water surfaces or else from moistened surfaces:

$$E_0 = \rho D (q_s - q), \quad (5.4)$$

where ρ represents air density; D is the integral coefficient of diffusion; q_s is the coefficient of humidity of air saturated by water vapour for a given temperature of the surface from which evaporation takes place; and q is specific air humidity at the level at which meteorological measurements are being taken.

In order to determine the temperature of the surface at which evaporation takes place and which determines the magnitude of q_s the heat balance equation may be employed (Budyko, 1971).

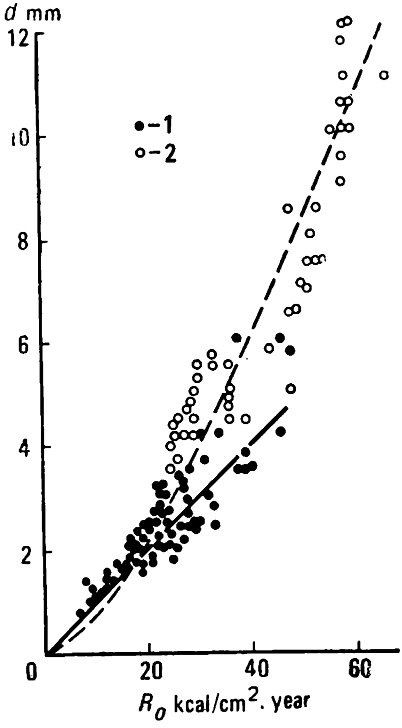

Fig. 22. A Comparison of Average Yearly Values of the Air Humidity Deficit with Values of the Radiation Balance: 1—Tundra, Forest, Forest-Steppe, Steppe; 2—Semi-desert, Desert

A comparison of magnitudes of E_0 calculated according to formula (5.4) with values of R_0/L for 44 locations in different climatic conditions is presented in Figure 21; it shows that calculations of potential evaporation according to formula (5.4) and proceeding from average yearly values of the radiation balance produce generally similar results, for which the correlation coefficient is 0.98.

The relation between potential evaporation and the air humidity deficit (whose values are often employed in calculating potential evaporation) may be seen in Figure 22. It presents average yearly values of the air humidity deficit as a function of the radiation balance for conditions at 187 locations in different natural zones within the USSR. Data on regions possessing dry climates (semi-deserts and deserts) are denoted by circles, while all other data relating to steppes, forest-steppes, forest zones and tundras are represented by dots. It may be seen from Figure 22 that for conditions of a more or less humid climate there is a relatively close relationship between values of the radiation balance and the air humidity deficit which approximates a direct proportionality. This relationship is violated in the case of dry climates.

Thus it is possible to determine potential evaporation from the radiation balance at the Earth's surface. The method for calculating potential evaporation from the air humidity deficit as well as a number of other empirical methods for determining potential evaporation yield less accurate results.

2. *A Periodic Law of Geographical Zonality*

The Table of Geographical Zonality. In order to study the influence of climatic factors on geographical zonality a world map of the radiation index of dryness was constructed (Budyko, 1955). This is shown in Figure 23. In constructing that map data on the radiation balance were employed that were calculated for moistened surfaces.

In constructing the map of indices of dryness values for the *R*/*Lr* index were calculated for 1600 locations evenly distributed over the surface of continents (except for mountainous regions which are represented by shaded areas).

It may be seen from that map that the distribution of isolines of radiation indices of dryness on continents is characterized by substantial changes in value on each

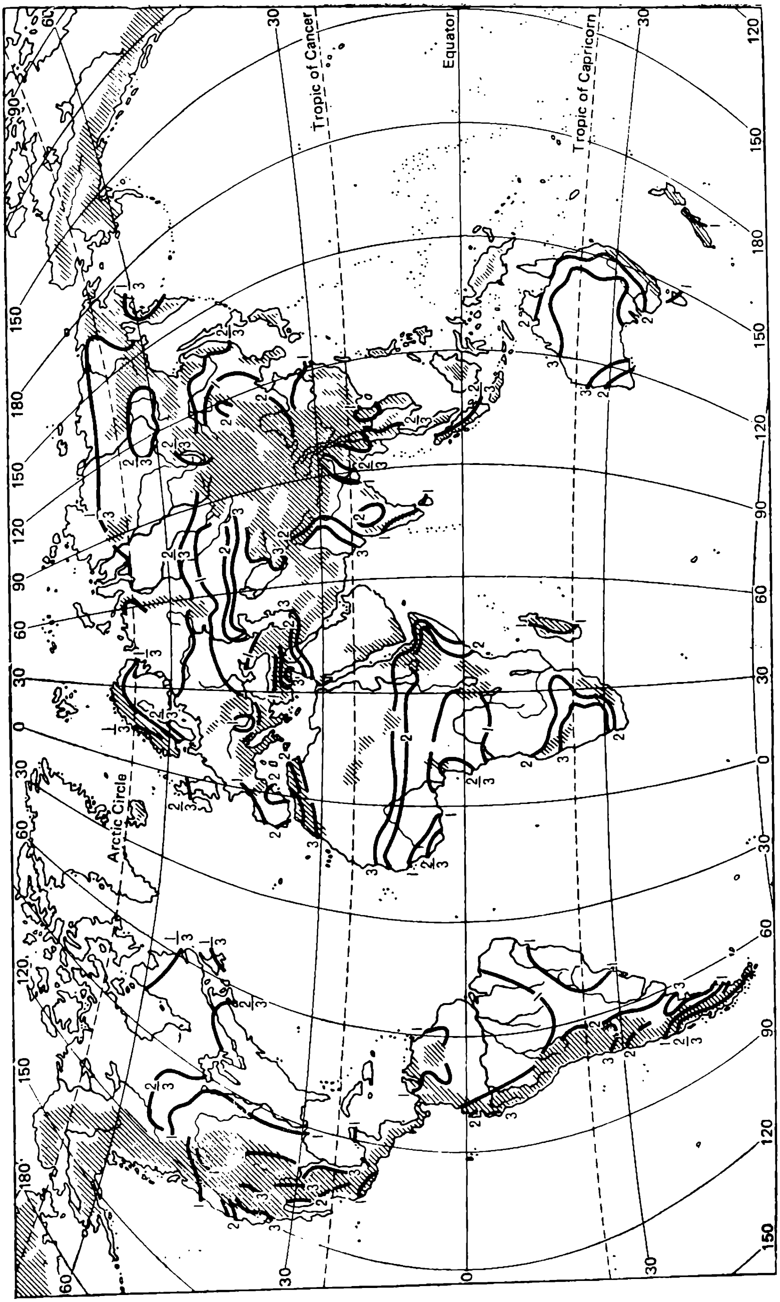

Fig. 23. Radiation Index of Dryness

continent. The most humid conditions (corresponding to the lowest values of the index of dryness) are largely found at high latitudes where the potential evaporation is very low. The largest values of the index of dryness are found in semi-desert and desert areas.

A comparison of the map of the radiation index of dryness with geobotanical and soil maps confirms that the position of isolines of the index of dryness conforms well to the distribution of major geographical zones.

The lowest values of the index of dryness (less than $^1/_3$) correspond to tundras, values between $^1/_3$ and 1 to the forest zone, from 1 to 2 to the steppe zone, more than 2 to the semi-desert zone, and more than 3 to the desert zone.

Neglecting for the moment a number of specific conclusions that may be derived from a comparison of a map of the index of dryness with geobotanical and soil maps let us emphasize the major point, namely that the radiation index of dryness which determines the relative value of components of the water and heat balances accords well with the boundaries of major natural zones. It is clear, moreover, that within particular zones at various latitudes one observes substantial differences in conditions determining the development of natural processes. These differences derive from the fact that the energy base of natural processes that may be characterized through the magnitude of the radiation balance R differs at various latitudes.

Accordingly, while it may be possible to make use of the single parameter R/Lr (which determines the relative values of components of the heat and water balances) in characterizing general zonal conditions of natural processes, a characterization of the absolute values of the intensity of natural processes requires the use of two parameters, namely R/Lr and R, which determine the absolute magnitudes of elements of the heat and water balances.

The relation of geobotanical conditions with the parameters R/Lr and R may be represented as a graph with R and R/Lr plotted along its axes and on which the major geobotanical zones are divided by straight lines. A schematic form of such a graph is presented in Figure 24.

It should be noted that the ordinate represents values of the radiation balance for the actual state at the Earth's surface, which differ from values of the radiation balance for moistened surfaces. The continuous line on the graph bounds the region of actually occurring values of R and R/Lr (except for mountainous regions), within whose ranges specific values of the parameter R/Lr, which are shown as vertical lines, delimit the major geobotanical zones: tundra, forest, steppe, semi-desert, desert. Large changes in values of the radiation balance for the forest zone and somewhat smaller changes for the steppe zone correspond to perceptible geobotanical changes within these zones in spite of the common features of their plant covers.

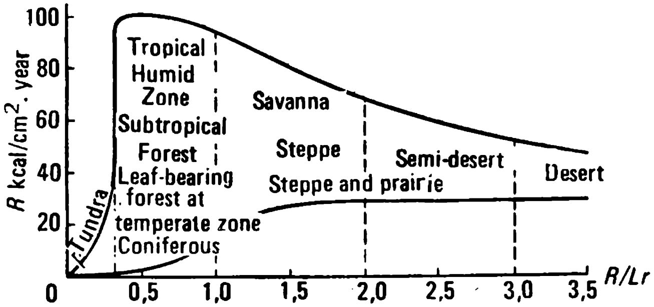

Fig. 24. A Graph Representing Geobotanical Zones

The close relation of the zonality of plant covers with the zonality of soils allow for conclusions concerning the relation of botanical zones to specific values of parameters R and R/Lr to be fully applied to soil zones as well.

This makes it possible to establish that, as values of the parameter R/Lr increase, soil types change in the following sequence: (*a*) tundra soils; (*b*) podzols, brown forest soils, yellow soils, red soils, and laterite soils (the diversity of types of soil within that group corresponds to changes in the parameter R within broad ranges); (*c*) chernozems and black soils of savanna regions; (*d*) chestnut brown soils; and (*e*) grey soils. The general relation of soil zonality to climatic indices R/Lr and R may be represented as a graph that is similar to the one for botanical zones in Figure 24.

The relation of the zonal structure of the hydrological regime of land to parameters R and R/Lr may be established in both quantitative and qualitative terms.

It follows from the linkage equation that to each gradation in value of the parameter R/Lr there corresponds a specific gradation in the value of the run-off coefficient. As a result, for tundra conditions, where $R/Lr < 1/3$, the run-off coefficient must be more than 0.7; for the forest zone possessing coefficients $1/3 < R/Lr < 1$ the run-off coefficient must have values ranging from 0.3 to 0.7; in the steppe zone (where $1 < R/Lr < 2$) it ranges from 0.1 to 0.3; and in the semi-desert and desert zones it is less than 0.1. Observed data confirm this regularity.

Thus a consideration of the influence of energy factors makes it possible to explain zonal changes in run-off coefficients in quantitative terms. The absolute values of total run-off are determined by two parameters, namely R and R/Lr. The graph in Figure 25, which is similar to Figure 24, represents the distribution of yearly total run-off. It characterizes the absolute values of run-off in different geographical zones.

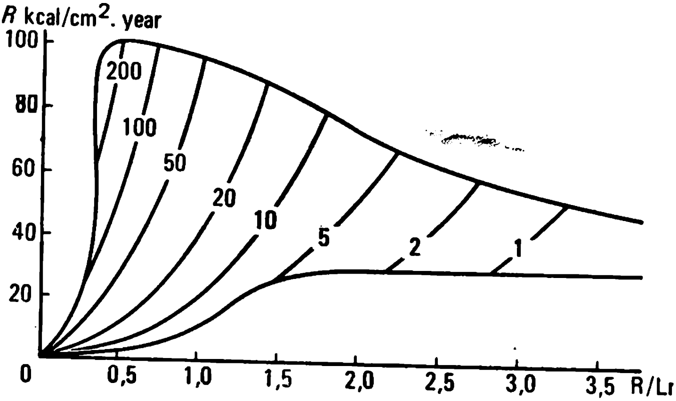

Fig. 25. Run-off (cm/year) under Different Climatic Conditions

It should be noted that data on total precipitation that are employed in these calculations do not reflect corrections that were proposed in recent years for determining precipitation normals found with the help of available measuring instruments. Such corrections result in increased values of total precipitation that are especially noticeable in regions receiving large volumes of solid precipitation.

While the use of more accurate precipitation normals introduces certain changes in the characteristic values of moisture indices, this does not alter the general patterns of the relation of moisture indices to geographical zones. In a number of cases the corrected total values for precipitation produce an improvement in the correspondence between moisture indices and boundaries of geographical zones.

It follows that a close relation exists between geographical zones and two climatic parameters, namely the radiation balance of the Earth's surface and the radiation index of dryness. This relation was examined in studies of Grigoriyev and the present author (1956, 1962), in which this regularity was named the Periodic Law of Geographical Zonality.

More specifically it was noted that within each latitudinal belt one finds a definite correspondence between the boundaries of natural zones and the isolines for particular values of the radiation index of dryness. Within the different latitudinal belts, moreover, one finds that natural zones that are similar in a number of important characteristics correspond to similar values of the radiation index of dryness. For example, in passing from one latitudinal belt to another (which leads to a change in values of radiation balance) one finds in a number of natural zones that are defined by changes in moisture conditions elements of continued similarity as well as differences, attributable to changes in thermal energy conditions.

Table 11 represents the corresponding overall structure It is a further development of the diagram presented in Figure 24. It should be noted that in compiling the table of geographical zonality use was made of the values of the radiation balance relating to the conditions of moistened surface. This is one of the reasons for data of the table differing from those given in Figure 24. Other differences result from the use, in compiling the table of geographical zonality, of more detailed data on precipitation and plant cover in various regions. The table leads to the following conclusions concerning the patterns of the geographical zonality of lowland regions. To each section in the table characterizing gradations of moisture conditions correspond specific values of coefficients of river run-offs. At the same time the yearly run-off normal in each section increases as radiation balance increases, i.e., as the latitude decreases (this does not apply to desert areas where the run-off at low latitudes is close to zero).

Table 11

TABLE OF GEOGRAPHICAL ZONES

Source of heat energy — radiation balance	Moisture conditions — radiation index of dryness								
	Less than 0 (highly excessive moistening)	Between 0 and 1					Between 1 and 2 (moderately insufficient moistening)	**Between 2 and 3 (insufficient moistening)**	Greater than 3 (highly insufficient moistening)
		Excessive moistening				Optimal moistening $^{4}/_{5}$-1			
		0-$^{1}/_{5}$	$^{1}/_{5}$-$^{2}/_{5}$	$^{2}/_{5}$-$^{3}/_{5}$	$^{3}/_{5}$-$^{4}/_{5}$				
Less than 0 (high latitudes)	**I** **Eternal snow**	—	—	—	—	—	—	—	—
Between 0 and 50 kcal/cm²·year (South Arctic, subarctic and moderate latitudes)	—	IIa Arctic desert	IIb Tundra (with islands of sparse forests in the south), bow forests in swampy areas	IIc Northern and middle taiga	IId Southern taiga and mixed forests	IIe Deciduous forests and forest-steppe	III Steppes	IV Semi-deserts in temperate belts	V Deserts in temperate belts
From 50 to 75 kcal/cm²·year (subtropical latitudes)	--	—	VIa Regions of subtropical gemigilei with substantial swamp areas	VIb Subtropical rain forests			VII Hard-leaved subtropical forests and brush, deciduous forests	VIII Subtropical semi-deserts	IX Subtropical deserts
More than 75 kcal/cm²·year (tropical latitudes)	—	—	Xa Regions in which equatorial forest swamps prevail	Xb High humidity equatorial forests (major swamps)	Xc Moderate humidity equatorial forests (moderate swamps)	Xd Equatorial forests, passing to light tropical and deciduous savannas	XI Arid savannas, deciduous forests	XII Semi-desert savannas (tropical semi-desert)	XIII Tropical deserts

Similar types of plant cover correspond to each section in the table. In conditions of excessive moisture (the third section) one finds a prevalence of forests at all latitudes, except for areas with substantial excess of moisture (the index of dryness less than $^2/_5$). In such cases forests are replaced by tundras at higher latitudes and by swamps at lower latitudes (since the index of dryness less than $^2/_5$ over vast territories is found only at relatively high latitudes, swampy areas at low latitudes cannot be viewed as autonomous geographical zones).

Each gradation in moistening is characterized not only by a definite type of plant cover, but also by a specific value of its productivity. In the joint studies of Grigoriyev and the present author mentioned earlier it was assumed that the productivity of natural vegetation increases as moisture conditions approach optimal ones (for a given radiation balance) and that for given moisture conditions productivity increases as the radiation balance increases. (The problem of the productivity of plant covers in different geographical zones is examined more closely in Chapter VI.)

To each section of the table corresponds a specific sequence of changes in soil types largely similar within each section. For example, the third section is characterized by the following sequence of soil types: tundras, podzols and brown forest soils, subtropical red soils and yellow soils, tropical podzol red soils and laterites; for the fourth section we find chernozems and dark chestnut soils, black soils and brown soils, weakly leached subtropical soils, red-brown tropical soils, etc.

Specific values of quantitative characteristics of the process of soil formation correspond to each such sequence.

A special place within the table is occupied by the region of eternal snow, whose characteristics are unique. It is described by a negative value of the radiation balance, a negative index of dryness and a practical absence of plant and soil covers. Similarly the subzone of Arctic deserts possesses negligible values of yearly radiation balance and a high humidity.

The patterns in the Table of Geographical Zones may be explained in terms of the following factors.

Because of the Earth's spherical shape its surface is divided into several latitudinal belts that differ in terms of radiation energy inflow to the Earth's surface. Within each of these belts (except for the region of eternal snow) one may find different moisture conditions, ranging from excessive to highly insufficient. It is not difficult to see that within different latitudinal belts geographical conditions with similar moisture indices will have a certain number of common characteristics (which are combined with differences that result from differing inflows of radiation energy). These common characteristics will be periodically repeated as we compare regions of two latitudinal belts in which humidity (or else dryness) increases.

Seasonal Changes in Climatic Factors of Geographical Zonality. The Table of Geographical Zones establishes a relation between natural zones and average yearly values of components of the climatic regime. It should be kept in mind that usually these values result from an averaging of the fluctuations in elements of climate that are characteristic for particular seasons.

By considering the specific features of the climatic regimes in different seasons the influence of climatic factors on geographical zones may be represented more clearly (Grigoriyev, Budyko, 1962).

In specifying the climatic conditions of particular seasons let us identify the following major types of climatic regime:

1. The Arctic climatic regime that is characterized by the snow cover, negative air temperatures and negative values of the radiation balance, or else values close to zero.
2. The climatic regime characteristic of tundras, with average monthly temperatures that range from zero to 10°C and a small positive radiation balance.
3. The climatic regime characteristic of forest zones, with average monthly temperatures of more than 10°C and a positive radiation balance, as well as sufficient moisture, when evaporation exceeds one half of the potential evaporation.
4. The climatic regime of arid zones (steppes and dry savannas), with a positive radiation balance and an evaporation that ranges from $^1/_{10}$ to $^1/_2$ of potential evaporation.
5. The climatic regime of deserts, with a positive radiation balance and an evaporation less than $^1/_{10}$ of potential evaporation.

If any of these types of climatic regime were observed throughout an entire year there would be a correspondence of the type of climatic conditions in each month with that of the natural zone. But in most geographical regions several types of climatic regime replace each other during the year. For example, four types of climatic regime replace each other during the year in the steppe zone of the European part of the USSR: during the winter period there is an Arctic regime; at the beginning of spring there is a regime characteristic of tundras; this is replaced by a climatic regime characteristic of the forest zone; and during the summer there is a climatic regime characteristic of the steppe zone. In that case only the last of the seasonal types of climatic regime corresponds to the type of geographical zone.

A general description of changes in climatic conditions during the course of a year at various latitudes of the Northern Hemisphere is provided in Figure 26A. This relates to the lowland regions in the zone between 20° and 40° East longitude which are located in Europe and Africa. Figure 26B represents changes in climatic conditions for the eastern outskirts of Asia, while Figure 26C represents similar changes for eastern regions of North America and for South America.

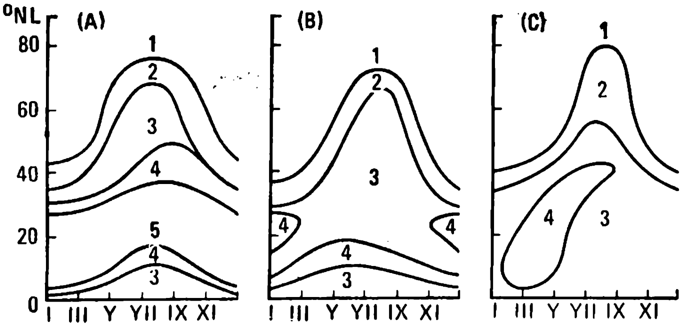

Fig. 26. Seasonal Changes in Climatic Factors Determining Geographical Zonality (1-5 are climatic regimes)

These figures reflect substantial differences in seasonal changes of climatic regimes at various longitudes, which are especially felt at middle and low latitudes where types of regime depend on moisture conditions.

In Europe and Africa a regime of low humidity exists within a wide latitudinal belt that shifts to the north in the summer and to the south in the winter. This corresponds to the most favourable moisture conditions in the subtropics in the winter and spring and at tropical latitudes in the summer.

In East Asia and North America the structure of the moisture regime differs considerably from the first scheme because of substantial differences in circulation processes.

It may be seen in Figure 26 that in most regions and at all longitudes one finds either insufficient moisture (regimes *4* and *5*) or else insufficient heat (regimes *1* and *2*) during a major part of the year. It is only in the narrow band close to the Equator that one finds conditions corresponding to regime *3* throughout the entire year.

Evidently under regimes *2* and *4*, and especially under regimes *1* and *5*, the productivity of the natural plant cover is reduced. One also finds the following regularity: under conditions of insufficient heat the type of geographical zone is determined by the climatic regime of the period in which the productivity of the natural plant cover is greatest, even if that period is relatively short. Thus, for example, the zonal landscape of the tundra is determined by conditions of the warm season, which may last no longer than $^1/_5$ to $^1/_4$ of the year. In such cases the climatic regime of the cold season, which extends over most of the year, does not determine the landscape's zonal character.

A similar regularity is usually also found when there is insufficient moisture and the most humid period of the year plays a determining role in establishing the type of zone, even though it may be shorter than the period with insufficient moisture.

It should be noted, however, that in natural zones with insufficient moisture there may be short periods with either sufficient or excessive moisture that do not bring about the development of the typical forest growth, which requires more or less prolonged soil moistening.

These regularities, which relate types of geographical zones to seasonal changes in the climatic factors that determine zonality, complement the conception of the influence of climatic factors on geographical zonality that has been described earlier.

The conception of a periodic law of geographical zonality formulated in 1956 is widely used in many current studies. In particular, it is used in order to clarify the patterns of distribution of plant covers (Armand, 1967; Bazilevich, Rodin and Rozov, 1970, et al.). In recent years it has been shown that it may be applied to study the zonality of geomorphological processes (Budyko, Kotlyakov and Meshcheryakov, 1970).

The significance of the periodic law for studying phenomena of geographical zonality has been examined in a monograph by S. V. Kalesnik (1970).

VI

The Circulation of Organic Matter

1. *Activities of Autotrophic Plants*

The Physical Mechanism of Photosynthesis. The organic matter in nature is produced as a result of the activity of autotrophic plants, which are the only group of organisms able to synthesize organic matter from minerals. There are two ways in which organic matter may be created: first, through the utilization of radiation energy (photosynthesis), and secondly, through chemical synthesis. But only the first of these is important for the production of biomass, since, while chemical synthesis is important for the circulation of nitrogen and for certain other processes, it produces a relatively small quantity of organic matter. Plants on land that are able to engage in photosynthesis use carbon dioxide from the air in order to construct organic matter. They also use water and a relatively small quantity of mineral substances from the soil, small as compared with the total mass of synthesized substance. Aside from carbon dioxide from the air plants may also receive some quantity of carbon dioxide from the soil, but it is the first of these sources that is important. A certain part of the energy of short-wave solar radiation is expended on the process of photosynthesis within the plant cover.

The principal content of the overall reaction produced by photosynthesis may be represented in a simplified way in terms of the following equation:

$$6CO_2 + 6H_2O = C_6H_{12}O_6 + 6O_2.$$

Many experimental studies have shown that the "efficiency" of plants engaged in photosynthesis (i.e., the ratio of energy expenditure on the synthesis of biomass to the total quantity of absorbed solar energy) is very modest, and usually does not exceed 0.1-1 per cent. Under favourable conditions its magnitude increases to several per cent (Nichiporovich, 1973).

It should also be noted that a large number of experimental data have shown that plants expend water resources very uneconomically. The productivity of transpiration (the ratio of the increment in weight of dry plant mass to expenditure of water on transpiration over a given time interval) usually ranges from $^1/_{200}$ to $^1/_{1000}$ (most often it is roughly $^1/_{300}$). The view has long been expressed that such an abundant transpiration does not follow from the plants' physiological needs and is largely a useless expenditure of water (N. Maximov, 1926, 1944).

These two fundamental facts indicate that under natural conditions the plant cover assimilates only a negligible part of available energy and water resources. In studying the relation of plant productivity to climatic factors it is obviously quite important to establish the causes of such a substantial limitation on the use of natural resources in the synthesis of biomass.

In order to consider this more closely let us turn to an analysis of the physical mechanism of assimilation of carbon dioxide, and also of transpiration (Budyko, 1949, 1956).

The organ of assimilation of a plant engaged in photosynthesis is its leaf. This is a casing of dense cuticular tissue with a large number of small apertures—stomata that may open and close. This casing contains a large area of chloroplasts containing grains of chlorophyll. The surface of chloroplasts is linked to atmospheric air through intercellular elements and stomata.

It is important to note that in order for photosynthesis to proceed the surface of chloroplasts must be maintained in a moist state, since carbon dioxide can be assimilated only in the form of a solution. As a result, relative air humidity in intercellular elements is large and usually exceeds substantially relative air humidity in the atmosphere. Thus the diffusion of carbon dioxide into a leaf with open stomata is inevitably accompanied by a diffusion of water vapour into the opposite direction, i.e., by a transpiration of the plant.

When the plant cover is relatively dense the following equation may be employed to express the heat balance for the layer within which the plant cover exists:

$$R = LE + P + lA, \tag{6.1}$$

where R is the radiation balance, E—the evaporation (transpiration), L—the latent heat of evaporation, P—the turbulent flow of heat from the Earth's surface to the atmosphere, A—assimilation, and l—the expenditure of heat needed to assimilate a unit mass of CO_2.

If we consider that for a sufficiently dense plant cover evaporation is largely determined by the process of transpiration, we will find that the three members on the right-hand side of the equation (6.1), namely evaporation, turbulent heat flow, and assimilation—depend on diffusion processes possessing similar mechanisms.

In the course of plant transpiration water vapour is diffused through intercellular elements and stomata from moist walls of a leaf's internal surface to its external surface, and the concentration of water vapour changes from q_s (the average concentration in air at the leaf's internal surface) to q_o (the average concentration at the leaf's external surface).

The rate of evaporation will then be:

$$E = \rho D' (q_s - q_o), \tag{6.2}$$

where ρ is air density and D'—the integral coefficient of diffusion for water vapour as it moves from the leaf's internal surface to its external surface. It is evident that the magnitude D' depends on the leaf's morphology (the number and dimensions of the stomata, the leaf's thickness, the density of cuticle, etc.).

A second stage in the diffusion of water vapour is associated with changes in the concentration of water vapour from q_o (at the leaf's surface) to q—the concentration outside the plant cover. That stage may be represented through the equation:

$$E = \rho D'' (q_o - q), \tag{6.3}$$

where D'' is the integral coefficient of external diffusion, which largely depends on the intensity of turbulent exchange.

Removing q_o from formulae (6.2) and (6.3), we find

$$E = \frac{\rho (q_s - q)}{\frac{1}{D'} + \frac{1}{D''}}. \tag{6.4}$$

The diffusion of carbon dioxide from the air outside the plant cover to the absorbing surface within the leaf may also be divided into two stages.

On the one hand, the rate of diffusion (which is equal to the rate of assimilation) is equal to:

$$A = \rho D'_c (c_0 - c_1), \tag{6.5}$$

where D'_c is the effective coefficient of diffusion through intercellular elements and stomata for carbon dioxide moving from the external surface of a leaf to the surface of parenchyma, c_0—the average concentration of carbon dioxide in the air at the leaf's external surface, and c_1—the average concentration of carbon dioxide in the air at the surface of parenchyma cells.

On the other hand, the rate of assimilation is

$$A = \rho D''_c (c - c_0), \tag{6.6}$$

where c is the concentration of carbon dioxide outside the plant cover, and D''_c—a magnitude similar to D''.

It follows from (6.5) and (6.6) that

$$A = \frac{\rho (c - c_1)}{\frac{1}{D'_c} + \frac{1}{D''_c}}. \tag{6.7}$$

The ratio of the rate of assimilation to the rate of evaporation is found from (6.4) and (6.7):

$$\frac{A}{E} = a \frac{c - c_1}{q_s - q}. \tag{6.8}$$

In this connection it should be noted that the coefficient of proportionality in that formula

$$a = \frac{D'_c D''_c (D' + D'')}{D' D'' (D'_c + D''_c)}$$

may change only within a relatively narrow range. Since the process of diffusion of water vapour and carbon dioxide from the external surface of a leaf and a free flow of air is determined by turbulent mixing, it is evident that the coefficients D'' and D''_c have equal values. It is natural to consider that the diffusion of water vapour and of carbon dioxide from the walls of a parenchyma to the external surface of a leaf is of a molecular nature. In that case the ratio D_c/D' must be equal to the ratio of the coefficient of molecular diffusion of carbon dioxide in the air to the coefficient of molecular diffusion of water vapour in the air, i.e., it will be roughly 0.64.

This makes it possible to conclude that depending on the ratio of the values of D' and D'' the coefficient a may change only between the values of 0.64 and 1, approaching one in cases when the leading role is played by external diffusion (a thin leaf with a large number of open stomata) and declining to 0.64 when the leading role is played by internal diffusion (a thick leaf with poor ventilation). If one takes that under average conditions the rate of evaporation from a leaf is approximately 50 per cent of the rate of evaporation from an equally moist surface, it is not difficult to calculate that for such a condition the value of coefficient a will be approximately 0.8.

Since the turbulent heat flow from the plant cover to the atmosphere is

$$P = \rho D'' c_p (T_w - T), \tag{6.9}$$

(where c_p is the air heat capacity, T—air temperature, and T_w—temperature at the leaf's surface), we find from (6.1), (6.4), (6.7), and (6.9) the formula for the rate of assimilation which will be:

$$A = \frac{R(c - c_1)}{\frac{L}{a}(q_s - q) + l(c - c_1) + b(T_w - T)}, \tag{6.10}$$

where $b = \frac{c_p}{a}\left(1 + \frac{D''}{D'}\right)$.

It follows from formula (6.10) that the rate of assimilation depends on the difference in concentrations of carbon dioxide outside the plant cover and at the internal surface of the leaf $(c - c_1)$ and that it increases as that difference increases. Since the concentration of carbon dioxide outside the plant cover is relatively constant, it follows that the rate of assimilation will increase as c_1 declines, i.e., as the absorption of CO_2 by the protoplasm becomes more intense.

When "physiological absorption" proceeds at an appreciable rate and the general rate of assimilation is limited not by physiological processes but by the diffusion flow of carbon dioxide, the value of c_1 must be substantially less than c and the formula for assimilation becomes:

$$A = \frac{Rc}{\frac{L}{a}(q_s - q) + lc + b(T_w - T)}. \tag{6.11}$$

With the help of formula (6.11) it is possible to determine the quantity of solar energy that may be assimilated by a plant when the utilization of carbon dioxide from the air is most effective.

Since c has an average value of $0.46 \cdot 10^{-3}$, $L = 600$ cal/g, $a = 0.8$, $l = 2{,}500$ cal/g, and $b = 0.62$ cal/g·deg, we find that

$$lA = \frac{1.2R}{770(q_s - q) + 1.2 + 0.62(T_w - T)}. \tag{6.12}$$

It should be noted that for conditions of photosynthesis that are optimal, when $c_1 \ll c$, and the leaf's internal surface is sufficiently moist, the value of q_s must approach the concentration of saturating water vapour at the leaf's temperature.

The average daily difference between the temperature of leaves and that of the air during the summer at middle latitudes is of the order of several degrees. Assuming that this difference is 5°C, and assuming as well that the average daily relative air humidity in the summer is close to 50 per cent and average air temperature is 20°C, we will obtain values of $q_s = 2.0 \times 10^{-2}$ (for a leaf temperature of 25°C) and $q = 0.7 \times 10^{-2}$. Substituting these values into formula (6.12) we find that $lA = 0.08\, R$, i.e., that the expenditure of energy on assimilation in average climatic conditions for middle latitudes may reach 8 per cent of the radiation balance.

Available data make it possible to establish that over the larger area of the European part of the USSR the radiation balance in the summer is 55-60 per cent of the total radiation arriving during that season. Substituting this value we will find that for average summer conditions at middle latitudes the natural plant cover, absorbing carbon dioxide most actively, may use approximately 5 per cent of inflowing solar radiation. That theoretically calculated estimate agrees well with empirical data relating to the "efficiency" of photosynthesis under the most favourable conditions (Nichiporovich, 1973).

Yet available empirical data indicate that such relatively high coefficients of utilization of solar energy are found only in individual cases, while the average ratio of energy expenditure on photosynthesis of natural plant covers to inflowing solar energy is usually of the order of 0.5 per cent (see below).

It should therefore be concluded that in average conditions the relation $c_1 \ll c$ does not hold, and that on the average the difference $c - c_1$ is approximately 10 per cent of the value of c. In other words, a comparison of the proposed theoretical diffusion scheme with actual data makes it possible to establish that plants usually use only a small part (of the order of 10 per cent) of the possible diffusion inflow of carbon dioxide.

It may be noted that this conclusion accords with the findings of a number of experiments carried out by V. N. Lyubimenko and other authors who have relied on physiological studies in establishing that leaves do not make full use of the possible diffusion of carbon dioxide. On the basis of these observations Lyubimenko (1935) concludes that "under natural conditions the production of dry matter is limited not so much by the small content of CO_2 in the atmosphere as by the insufficiently rapid operation of the enzymatic apparatus that governs the flow of matter to be assimilated and its assimilation".

The above scheme of the physical mechanism of assimilation and transpiration makes it possible to explain why plants expend water in such an uneconomic way in the course of their development, i.e., why the observed values of the productivity of transpiration are so small.

If one substitutes into formula (6.8) the average estimates for q_s and q that have been found for summer daytime conditions and assuming in accordance with the resulting estimates that $c - c_1 = 0.1c$, then the ratio A/E will be approximately $^1/_{360}$. This means that in assimilating one gram of carbon dioxide a plant loses approximately 360 grams of water through transpiration.

It should be noted that while the ratio A/E does not coincide exactly with the productivity of transpiration,* we may state that these values differ relatively little. Thus the estimate that has been found provides a theoretical explanation for the observed orders of magnitudes of productivity of transpiration.

The manner in which formula (6.8) explains the considerable measure of dependence of the productivity of transpiration on the air humidity deficit deserves close attention. That dependence is frequently noted by various researchers who have observed a perceptible increase in the productivity of transpiration as the air humidity deficit is reduced, both in greenhouse conditions and in passing from dryer climates to more humid ones.

The principal conclusion may be formulated as follows.

A plant usually employs a very small part of available natural energy and water resources. In fact, that part is small even when compared to the modest "efficiency" that may be achieved with a maximal possible diffusion of carbon dioxide from the air. The findings that were presented in this section indicate that when the carbon dioxide of the atmosphere is fully used plants are able to assimilate not less than 5 per cent of the received solar energy, and that under such conditions the productivity of transpiration should equal not several thousandths (as is usually the case) but several hundredths. Since one does not normally find such high indicators of effectiveness of photosynthesis in natural conditions it must be concluded that the actual productivity of plants is greatly limited by factors whose nature will be considered more closely below.

Photosynthesis Within the Layer of Plant Cover. The formulae that have been mentioned earlier relate to a single leaf or else a group of leaves that exist in similar environmental conditions.

In determining the total photosynthesis within a layer of plant cover it is necessary to consider differences between the meteorological factors of photosynthesis that exist at various levels of the plant cover.

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* In calculating the productivity of transpiration the incremen in dry organic matter is considered. This differs somewhat from the magnitude of assimilation because of the utilization by the plant of water and of mineral substances from the soil, and because of the loss of a part of dry matter during the plant's development on respiration and the decay of individual organs of the plant.

Such a problem was first formulated in the studies of M. Monsi and T. Saeki (1953), and J. Davidson and J. Philip (1958) that contained calculations of the total photosynthesis taking place in a plant cover, with due consideration of the influence of changes in the quantity of radiation from top to bottom, but without considering the dependence of photosynthesis on changes in the concentration of carbon dioxide. Subsequently a more general theory of photosynthesis in the plant cover was proposed that included a consideration of the influence on photosynthesis both of radiation factors and of changes in concentrations of carbon dioxide at various levels of the plant cover (Budyko, 1964; Budyko and Gandin, 1964, etc.; Budyko, Gandin and Yefimova, 1966). The findings of these studies are presented below.

Studies of photosynthesis at various levels of the plant cover must take into account physical regularities of meteorological processes in the layer of air within which the plants are found, i.e., within the layer of the plant cover. In studying physical processes within the layer of the plant cover it is appropriate to average values of meteorological elements at various levels along the horizontal line, which makes it possible to exclude the influence of individual plants on the meteorological regime.

By applying such averaging methods it is possible to establish that as distinct from the Earth surface air layer the average vertical flows of short-wave and long-wave radiation, heat and water vapour within the layer of plant covers and the momentum of the system depend substantially on height. More specifically, the quantity of short-wave radiation declines as one approaches the Earth's surface (due to the absorption and radiation scattering by the surface of plants). The resulting radiation flux (radiation balance) also declines in the same direction because of the screening effect of plants.

Similarly flows of water vapour and the momentum of the system change as well. In particular the flow of water vapour within the layer of the plant cover increases with height because of the influence of the plant's transpiration, while the momentum of the system declines as one moves downwards from the plant cover's upper boundary as a result of the inhibiting effects of plants on the movement of air.

This effect is associated with the reduction in turbulent exchange within the layer of the plant cover compared with higher air layers. The coefficient of turbulent exchange within that layer also declines as one approaches the Earth's surface.

An illustration of the dependence of the coefficient of turbulent exchange k and the photosynthetically active radiation Q on height H is presented in Figure 27. This is based on data produced by N. Yefimova (1968) for a field under wheat 1.2 meters high.

Let us note that the photosynthetically active radiation is a part of the general flux of short-wave radiation that is bounded by the wave lengths within which radiation may be employed for photosynthesis. Its value is approximately one-half of that of short-wave radiation.

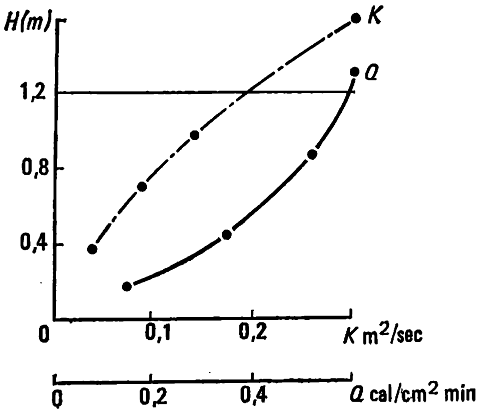

Fig. 27. Coefficient of Turbulent Exchange and Photosynthetically Active Radiation Within the Plant Cover Layer

In developing a theory of photosynthesis within the plant cover layer it is necessary to establish the quantitative relation between photosynthesis for a single leaf of a given plant and the external factors influencing it. Many laboratory studies have shown (Rabinovich, 1951) that for small values of the photosynthetically active radiation Q photosynthesis is proportional to the magnitude of radiation and is only slightly dependent on other external factors, i.e., that for

$$Q \to 0 \qquad A = \beta Q. \tag{6.13}$$

Such a condition corresponds to the ascending part of the light curve of photosynthesis.

When large amounts of radiation are involved, photosynthesis depends only slightly on incoming radiation (the plateau of the light curve) and depends substantially on the concentration of carbon dioxide, temperature, and other external factors.

Experimental studies have established that under conditions of "light saturation" photosynthesis is proportional to the concentration of carbon dioxide for the range of concentrations that are observed in natural conditions. In view of that conclusion we will assume that for

$$Q \to \infty \qquad A = \tau c, \tag{6.14}$$

where the coefficient of proportionality τ is a function of the temperature and of other external factors, but does not depend on radiation and on the concentration of carbon dioxide.

Apparently the values β and τ may be determined from experimentally established light curves of photosynthesis, moreover β is a tangent of the slope of the initial sector of the light curve, while τ corresponds to the height of the plateau of the light curve for a given concentration of carbon dioxide.

It should be noted that the coefficient τ describes conditions of diffusion of carbon dioxide that is assimilated by plants in the course of photosynthesis in the case of light saturation.

These conditions are met by a formula (Budyko, 1964), describing the light curve of photosynthesis:

$$A = \frac{\beta Q}{1 + \frac{\beta Q}{\tau c}}. \tag{6.15}$$

This represents a generalization of the hyperbolic formula proposed by Tamiya in 1951 (Monsi, Saeki, 1953).

The vertical flow of carbon dioxide within the layer of the plant cover at the level z is

$$A_z = \rho k \frac{dc}{dz}. \tag{6.16}$$

Change in that flow within the layer dz resulting from the assimilation of carbon dioxide by leaves will, according to (6.16), be as follows

$$dA_1 = \frac{\beta Q s dz}{1 + \frac{\beta Q}{\tau c}}, \tag{6.17}$$

where s represents the surface of leaves per unit volume.

Change in the photosynthetically active radiation with height within the layer of the plant cover are determined by the equation

$$dQ = \varphi(Q) s dz, \tag{6.18}$$

where the function $\varphi(Q)$ describes the dependence of the absorption of radiation within a layer with a thickness dz on the quantity of photosynthetically active radiation arriving at the upper boundary of that layer. Since at the upper boundary of a plant cover for $z=H$ the radiation flux is known ($Q=Q_H$), we find from (6.18) that

$$\int_Q^{Q_H} \frac{dQ}{\varphi(Q)} = \int_z^H s dz. \tag{6.19}$$

Solving equation (6.19) for Q we find a relation of type

$$Q = \psi\left(Q_H, \int_z^H s dz\right), \tag{6.20}$$

which describes the dependence of the photosynthetically active radiation on height. If we consider that

$$\varphi(Q) = \gamma Q, \tag{6.21}$$

that relation will have the form

$$Q = Q_H e^{-\gamma \int_z^H s dz}. \tag{6.22}$$

Making use of the relation (6.22) we obtain from formula (6.18):

$$dA_1 = \frac{\beta Q_H e^{-\gamma \int_z^H s dz}}{1 + \frac{\beta Q_H}{\tau c} e^{-\gamma \int_z^H s dz}} s dz. \tag{6.23}$$

Change in the flow A_2 produced by the respiration of the plants may be expressed in the form

$$dA_2 = -\varepsilon s dz, \qquad (6.24)$$

where ε is the coefficient of proportionality.

The general change in the vertical flow of carbon dioxide is determined by the relation

$$dA_z = dA_1 + dA_2. \qquad (6.25)$$

From these formulae we obtain the nonlinear differential equation

$$\frac{d}{dz}\left(\rho k \frac{dc}{dz}\right) = -\frac{\beta s Q_H e^{-\gamma \int_z^H s dz}}{1 + \frac{\beta Q_H}{\tau c} e^{-\gamma \int_z^H s dz}} - \varepsilon s. \qquad (6.26)$$

In order to solve equation (6.26) let us use the following boundary conditions:

$$\left(\rho k \frac{dc}{dz}\right)\Big|_{z=0} = -A_0, \qquad (6.27)$$

$$\left(\rho k \frac{dc}{dz}\right)\Big|_{z=H} = \rho D_H (c_\infty - c_H), \qquad (6.28)$$

where A_0 is the flow of carbon dioxide from the soil; c_∞ is the concentration of carbon dioxide in the air at the height z_1, where its dependence on the properties of the plant cover is small; c_H is the concentration of carbon dioxide at the level H; and D_H is the integral coefficient of turbulent diffusion within the air layer between levels H and z_1.

The first of these conditions expresses an equality between the inflow of carbon dioxide from the soil and the magnitude of the turbulent flow of carbon dioxide in the air at the soil's surface. The second condition corresponds to an equality between the turbulent flow of carbon dioxide arriving at the upper boundary of the plant cover with the turbulent flow arriving from the atmosphere into the plant cover layer.

From the balance condition for carbon dioxide in the space between leaves for the entire plant cover we find that

$$A = A_H + A_0 + \int_0^H \varepsilon s dz, \qquad (6.29)$$

where A is the total magnitude of assimilation; and A_H is the vertical flow of carbon dioxide at the level H.

A solution to equation (6.26) makes it possible to determine the value of A_H and then, with the help of formula (6.29), to find the magnitude of total assimilation. In this connection one should consider the dependence of k, s, and ε on height within the plant cover layer. This may be achieved with the help of available empirical data.

In the general case equation (6.26) may only be integrated in numerical terms.

Application of the theory referred to above makes it possible to study the dependence of photosynthesis within the plant cover layer on the magnitude of the index S (which is equal to the ratio of the surface of leaves to the area occupied by the plant cover) and on various external factors.

In recent years, aside from that theory of photosynthesis within the plant cover layer, other schemes for constructing numerical models of this process have also been proposed (Ross, 1964; Monteith, 1965; De Wit, 1965; Uchijima, 1966; Duncan, Loomis, Williams and Hanau, 1967; Tooming, 1974). They are all based on the general idea of a transition from photosynthesis within a single leaf to photosynthesis in a layer that is homogeneous horizontally but possesses different physical conditions at various heights. The differences between these schemes relate to different degrees of detail in constructing the corresponding numerical models; some of them are more general while others are more specific.

Let us consider some of the findings resulting from application of the theory that has been presented for studying the regularities of photosynthesis within the layer of plant cover (Gandin, Menzhulin and Usov, 1969). These relate to the special case in which the values of τ, s and ε do not depend on height, and in which the vertical profile of the coefficient of turbulent diffusion in the space between leaves is linear, i.e., $k = k_0 + k_1 z$, where $k_0 \ll k_1 H$.

Figure 28 represents the calculated dependence of assimilation A on the flow of photosynthetically active radiation at the upper boundary of the plant cover (Q_H) for various values of the index of leaf area $S = \int_0^H s dz$ and of the ratio τ/ρ.

This relation has been calculated for the following values of other parameters:

$A_0=\varepsilon=0$; $D_H=0.6$ cm/sec; $k_1=3$ cm/sec; $\gamma=1$; $\beta=8\cdot10^5$ g/cal; $k_1H/k_0=200$; $\rho=1.29\cdot10^{-3}$ g/cm^3; $c_\infty=0.46\cdot10^{-3}$ (this represents 0.03 of volume per cent).

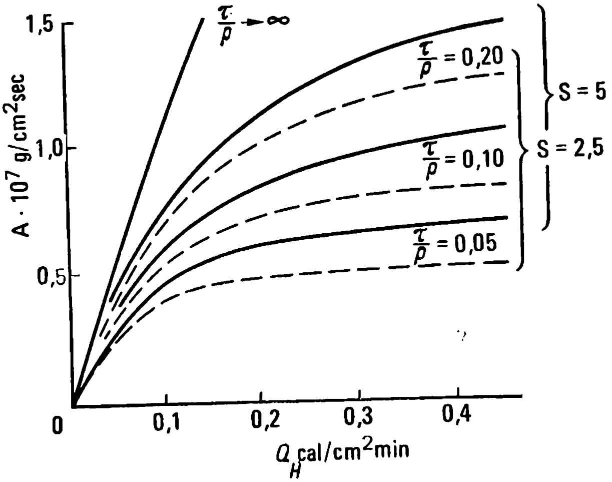

Fig. 28. Dependence of Assimilation on Radiation

It is not difficult to see that the lines on Figure 28 represent light curves. Unlike in the case of light curves corresponding to individual leaves, however, which have been studied in detail in terms of the physiology of plants (Rabinovich, 1951), they represent light curves corresponding to the plant cover as a whole derived in a theoretical manner. These curves show a strong dependence of total assimilation on the radiation flux for low levels of radiation, when assimilation is not very sensitive to other factors. For large values of radiation, on the other hand, the total assimilation is almost independent of it, and is determined by other factors, in particular by the temperature which influences the coefficient τ.

Figure 29 represents the influence of the inflow of carbon dioxide from the soil A_0 on total assimilation. Line *1* describes the actual dependence of A on A_0, while line *2* corresponds to the case when there is a full assimilation of carbon dioxide from the soil. It may be seen by comparing lines *1* and *2* that A increases as A_0 grows at a much lower rate than would occur if all additional carbon dioxide from the soil were expended on assimilation. It will be easy to understand this, if one considers that the flow of carbon dioxide from below first encounters leaves located in the shade

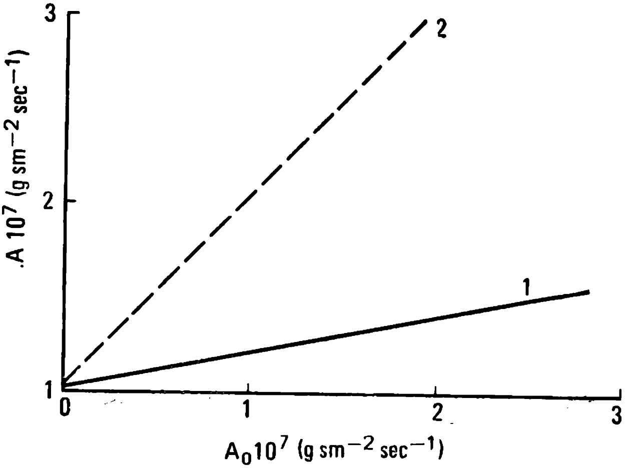

Fig. 29. Influence of Carbon Dioxide Flows from the Soil on Assimilation

and not able to engage in intensive photosynthesis because of insufficient radiation. As for the general increase in concentration produced by the inflow from below this is largely compensated by a reduced contribution of carbon dioxide from above.

As a result, the influence of the flow of carbon dioxide from the soil on assimilation is usually small by comparison with the flow of carbon dioxide arriving from the atmosphere. That conclusion accords with the findings of Monteith and other authors (Monteith, Szeicz and Yabuki, 1964) on the basis of experimental studies of the diffusion of carbon dioxide in plant covers.

Let us now turn to a consideration of the climatic factors that determine the productivity of plant cover *П*. We will consider that this productivity is equal to the difference between total assimilation and the expenditure of organic matter on respiration, neglecting losses of matter attributable to the atrophy of individual plant organs as the plant grows.

Since the productivity of plant cover depends substantially on the value of the index of leaf surface, and declines for both small and large values of that index (see below), we will calculate the value of its productivity for an optimal value of the leaf index, namely one that corresponds to the highest values of productivity.

The equations mentioned earlier show that in addition to parameters characterizing the properties of the plant cover itself the magnitude Π is also influenced by a considerable number of meteorological parameters. It therefore seems appropriate to identify the more important ones. It follows from the corresponding analysis that in conditions of sufficient moisture the influence of two factors affecting the productivity of plant cover is especially great, namely the level of photosynthetically active radiation and temperature. The value of the flux of radiation Q_H is represented in the equation. As for the temperature T, it influences the parameters τ and ε, which are employed in the calculation procedure.

Unfortunately, in spite of numerous studies in this area, we do not possess today sufficiently abundant and reliable quantitative data concerning the influence of temperature on parameters τ and ε for different types of plant. Accordingly, we have relied on the functions $\tau(T)$ and $\varepsilon(T)$ in a generalized form that is characteristic of certain grain crops.

With the help of these relations and of the equations noted earlier it is possible to calculate the dependence of the total productivity Π on the flux of photosynthetically active radiation at the upper boundary of the plant cover and on temperature. These calculations indicate that increases in the flux Q_H produce in all cases a growth in productivity Π. At the same time as temperature T increases the productivity Π first increases and then, after attaining a certain maximal value which depends on Q_H, begins to decline. This is a consequence both of the similar relation between τ and T, and of increased respiration as temperature increases.

The resulting calculations show that the influence of radiation and temperature on the productivity of plant cover is a complex one. In reality, moreover. radiation is always a factor whose value is a "minimum". This is associated with a constant insufficiency of radiation for lower layers of leaves of plant cover, which are partly in the shade. The influence of temperature on productivity, on the other hand, is different: increases in temperature above a certain threshold result in substantial declines in productivity.

Let us pause to consider the causes that may reduce the productivity of plants by comparison with the values derived from these calculations. These causes are associated with the influence of climatic, biological and soil factors.

The values for productivity that were calculated with the help of these methods relate to optimal moisture conditions, which in fact are found very infrequently throughout the entire period of the development of plants, even in the most favourable climatic regions.

The values of potential productivity refer to plants whose structure is optimal from the point of view of photosynthesis over the entire vegetation period, and whose photosynthesizing organs display a constant level of activity throughout that period. In reality such conditions are not met.

In addition, the possible productivity of plants corresponds to conditions in which they find a sufficient mineral nutrition. This, too, seldom occurs in practice, even in cases when adequate forms of agricultural technology are employed.

All these factors account for a substantial reduction in the productivity of both agricultural plants and of natural plant covers by comparison with their potential productivity, which is calculated through the procedure described above.

Accordingly, their potential productivity is a conventional indicator of productivity that describes the climatic resources of various geographical regions from the point of view of the creation of organic matter by autotrophic plants.

The water regime of soils plays a particularly large role in influencing the productivity of plants in real conditions. The findings of an empirical study of the dependence of the productivity of the plant cover on precipitation are cited below. The possibilities of applying methods of numerical modelling of the process of heat exchange and water exchange in plants in order to study the influence of water regimes on productivity deserve particular attention.

Studies of this type include the works of G. V. Menzhulin, who postulated that in conditions of insufficient water in the soil the leaf surface of plants declines, as does the stomata conductivity of leaf surfaces. With the help of a semi-empirical procedure for describing the corresponding regularities and with due account of the relation between transpiration and overall evaporation, Menzhulin calculated the influence of evaporation (and also of the associated soil moisture) on the productivity of plant cover (Menzhulin, 1976).

The Balance of Organic Matter in Plant Covers. Earlier it was noted that the diversity of types of plant cover in various regions of the world is largely explained by the influence of climatic conditions on the development of plant life.

Numerous studies concerned with the influence of climate on plant life have been largely based on empirical methods, and this reduces possibilities for identifying the nature of the causal mechanism through which climatic factors influence the properties of plant covers. At the present time it is possible to envisage studies of the dependence of properties of plant covers on climatic conditions on the basis of general physical regularities describing the influence of atmospheric factors on the vital processes of plants. Such an approach would make it possible to arrive at theoretically grounded estimates of average values of such characteristics of plant covers as their volume of biomass per unit surface for geographical zones corresponding to various climatic conditions.

In this connection let us consider the balance of organic matter in plant covers.

In analyzing the balance of organic matter in plant covers we will rely on the following relationship:

$$M = \mu S + m, \tag{6.30}$$

where M is the mass of organic matter contained in the plant cover per unit surface that the latter occupies; S is the leaf index; μ is the mass of leaves per unit of leaf surface; m is the mass of organic matter contained in all plant organs other than leaves per unit surface.

The productivity of plant covers per unit surface is determined through the formula

$$\Pi = A(S) - B(M), \tag{6.31}$$

where $A(S)$ is the photosynthesis within the layer of the plant cover, which depends on the value of the leaf index S; $B(M)$ is the expenditure of organic matter on respiration and on the atrophy of individual organs of plants, which depends on the mass of organic matter M.

It is reasonable to assume that the expenditure of organic matter on the respiration of plants and on their atrophy is proportional to the mass of leaves, i.e., is equal to $v\mu S$, where v is a coefficient of proportionality. In such cases

$$\Pi = A(S) - v\mu S - B(m). \qquad (6.32)$$

In order to establish the relation $A(S)$ one should make use of the physical theory of photosynthesis in a layer of plant cover that was described earlier.

In view of the nature of that relationship it may be established that the value $\frac{\partial A}{\partial S}$ is positive, declines monotonously as S increases, and approaches zero for large values of the leaf index. In that connection if $\left.\frac{\partial A}{\partial S}\right|_{S=0} > v\mu$, then there exists such a value of the leaf index S_0 at which productivity of the plant cover attains a maximal value. The value S_0 may be found from the condition

$$\left.\frac{\partial A}{\partial S}\right|_{s=s_0} = v\mu. \qquad (6.33)$$

It seems probable that as a result of a prolonged evolution of autotrophic plants directed at increasing the effectiveness of their photosynthetic activity, natural plant covers possessing sufficient moisture for transpiration will show values of the leaf index that more or less correspond to the optimal value S_0. In order to verify that proposition with the help of the calculation procedure described earlier, equation (6.33) was employed to determine the value S_0 for various climatic conditions. It was found that that value was relatively stable and declined slowly as the latitude increased and also that for leaf-bearing plants its average value was approximately 5.

Such an estimate accords satisfactorily with values of the leaf index obtained from observations of numerous natural plant covers in regions of sufficient moistening (*Foundations of Forest Biogeocenology*, 1964), and this confirms the assumption made earlier.

The development of perennial plants continues until they attain a natural maturity, i.e., a state in which the intake of organic matter in the plant cover is compensated by expenditures on respiration and atrophy, both of individual plant organs and of individual plants. In such cases the following condition holds true:

$$A(S) - v\mu S - B(m) = 0. \qquad (6.34)$$

Assuming that $S = S_0$ let us find the limiting value of organic mass M_0 that may be accumulated in a plant cover that has obtained a stationary state from the equation

$$A(S_0) = B(M_0). \qquad (6.35)$$

This equation may be employed for theoretical estimates of the quantity of biomass contained in typical plant covers in different natural zones of sufficient moisture.

Following the idea proposed by J. Davidson and J. Philip (1958), it is possible to construct an equation that will describe changes in the biomass of plant covers over time. This will make it possible to estimate the rate of accumulation of biomass as the plant cover approaches a state of natural maturity.

The form of that equation is similar to (6.32):

$$\frac{dM}{dt} = A(S) - v\mu S - B(m), \qquad (6.36)$$

where dM/dt is the rate of change of biomass over time.

We have employed equations (6.35) and (6.36) in order to estimate the value of reserves of biomass in plant covers in zones of sufficient moisture both in stationary states and when they change over time. Such calculations are made difficult by the insufficient comprehensiveness of available data on the intensity of the processes that are associated with the loss of biomass on respiration and on the atrophy of individual plant organs, needed for establishing v and B.

Nevertheless the use of values for these parameters that are available in the literature makes it possible to calculate through formula (6.35) the values of biomass reserves in forest at different latitudinal zones that accord satisfactorily with the data found in the study of Ye. N. Lavrenko, V. N. Andreyev, and V. L. Leontiyev (1955). Such calculations make it possible to establish which specific climatic factors influence changes in reserves of biomass during transitions from one type of plant cover to another.

In particular the formula indicates that declines in reserves of forest biomass that are associated with higher geographical latitudes are mostly caused by reductions in the duration of the vegetation period, while the influence of reduced air temperature during the vegetation period and of changes in radiation is less significant.

Calculations based on formula (6.36) make it possible to identify general patterns of changes in biomass during the development of perennial plant covers. These show increasing rates of accumulation of biomass during the initial years of existence of a plant cover, which are eventually replaced by a gradual decline in the rate of accumulation. When the rate of accumulation becomes small the plant cover approaches a state of natural maturity.

Such calculations have shown that formula (6.36) determines relationships of reserves of forest biomass to the age of the forests that accord well with data from corresponding observations.

These calculations may be compared with the well-known studies of G. F. Khilmi (1957, 1966), who has developed a method for calculating the growth of forest biomass on the basis of a dimensional theory and general physical conceptions. A comparison of the result of calculations based on equation (6.36) with corresponding calculations carried out by Khilmi shows that they produce a similar time pattern of biomass reserves in plant covers.

2. The Global Biological Cycle

The Productivity of Plant Covers. It has already been noted that the effectiveness of natural plant cover in accumulating the energy of solar radiation varies considerably, depending on natural conditions. The efficiency of solar energy in photosynthesis declines when there is a shortage of heat or moisture, and attains maximal values for optimal conditions of heat and water regimes. It also depends on the physical and chemical properties of soils.

Quantitative calculations relating to the utilization of solar energy by natural plant covers require information concerning the inflow of solar radiation and the productivity of plant cover. In studying the radiation regime at the

Earth's surface both maps of total radiation and maps of photosynthetically active radiation were prepared for various territories and for the world as a whole (N. A. Yefimova 1965, 1969, etc.).

In arriving at average values of the productivity of natural plant covers it is possible to employ data of various studies including empirical determinations of productivity in individual geographical zones. In recent years the studies of H. Lieth (1964-65), N. I. Bazilevich and L. Ye. Rodin (1967), as well as similar studies of other authors have led to the compiling of the first maps of plant productivity for various continents. While recognizing the great importance of these studies for further studies of the geographical patterns of the productivity of plant covers, it is also important to note that because of the insufficiency of initial data these maps were greatly simplified. In particular, on the maps prepared by Bazilevich and Rodin data on productivity are represented not as isolines but in terms of average values of productivity for different geobotanical zones. The use of such maps for studying the spatial distribution of the coefficient of solar energy utilization by plants is made difficult, for that productivity may change not only from zone to zone but also within individual zones.

In order to estimate the utilization of solar energy by plant covers a new map of the productivity of natural plant cover was compiled for the USSR (Budyko, Yefimova, 1968). The corresponding calculations reflected the fact that in conditions of sufficient moisture the productivity of plants depends on the radiation balance of the Earth's surface. The ratio of the radiation balance to the quantity of precipitation was employed as an index of moisture conditions.

A corresponding graph was plotted on which empirical values of the productivity of natural plant covers over a year (yearly increments in organic matter both above ground and below ground) were compared with values of the radiation balance and of precipitation. That graph is shown in Figure 30, in which the vertical axis represents values of the productivity of dry organic substances (centners per hectare, c/ha) over a year, while the horizontal axis represents the ratio of yearly values of the radiation balance R to the quantity of heat Lr, required for a full evaporation of total yearly precipitation (the radiation index of dryness). The curves represented on that graph correspond to different values of the radiation balance (kcal/cm^2·year).

While all data relating to the productivity of plant cover available in the literature were employed in constructing that graph, they were very few in the case of low latitudes (i.e. for large values of R), and this reduces the accuracy of the graph for the corresponding regions. For middle latitudes, however, the relationship represented on that graph should be more reliable.

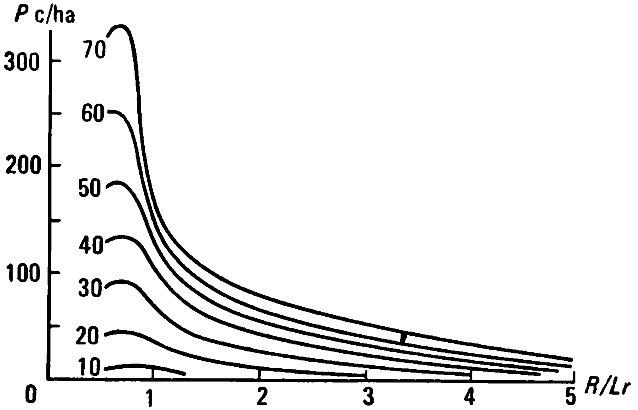

Fig. 30. Dependence of the Productivity of Natural Plant Covers on Meteorological Factors

The system of isolines on that graph agrees well with the law of geographical zonality described in Chapter V. It may be seen that changes in productivity attributable to the index of dryness at various levels of the radiation balance are largely similar. Similar relations between productivity and climatic factors were found in the studies of Bazilevich and Rodin (1969) and of A. V. Drozdov (1969). Some of the specific features of these relations call for additional studies. This includes the question of the dependence of maximal values of productivity for given magnitudes of the radiation balance on the radiation index of dryness. It may be seen from Figure 30 that it is not very easy to determine the precise position of maximal values for the corresponding curves. In the study of D. L. Armand (1967) it was suggested that the corresponding maxima occur in climatic conditions of high and middle latitudes, while in the tropics the productivity increases without constraints as the index of dryness is reduced. Bazilevich, Rodin and N. N. Rozov (1970) have concluded that such maxima exist at all latitudes and that as the latitude increases so does the corresponding value of the radiation index of dryness.

In compiling a map of productivity based on Figure 30 and on data concerning the radiation balance and the precipitation regime, calculations of productivity were carried out for 200 localities on the territory of the USSR. This was followed by the compiling of the map itself (Figure 31). It may be seen from this map that values of the yearly increment of biomass in the natural plant cover range from less than 20 c/ha on islands and the Siberian coast of the Arctic Ocean to values of more than 200 c/ha on the Black Sea coast of the Caucasus. In zones that are sufficiently humid, moreover, one finds that productivity increases as one moves from north to south in accordance with increases in the inflow of solar energy and the duration of the vegetation period. A reduction in the productivity of plant cover is found in regions of insufficient humidity, while in the deserts of Central Asia productivity declines to 20 c/ha per year.

A comparison of the values for the yearly productivity of natural plant covers on that map with the data from the maps prepared by Bazilevich and Rodin generally shows a satisfactory degree of correspondence between the characteristics of yearly increments of biomass for various geobotanic zones. Some of the differences between these maps in zones of dry steppes, deserts, and partly in regions of Eastern Siberia may be explained by an insufficient volume of empirical data for these territories. With the help of the map representing the productivity of plant cover and the map of influx of photosynthetically active radiation during the vegetation period (Yefimova, 1965) it is possible to calculate the share of incoming radiation energy that is assimilated in the organic matter formed during the year. In carrying out these calculations use was made of an approximate value of the caloricity of dry organic matter of plants, namely 4,000 cal/g.

Figure 32 represents a map of the distribution of energy contained in the yearly increment of organic mass of the plant cover that is based on these calculations. The corresponding values are expressed in terms of the percentage of energy of photosynthetically active radiation arriving du-

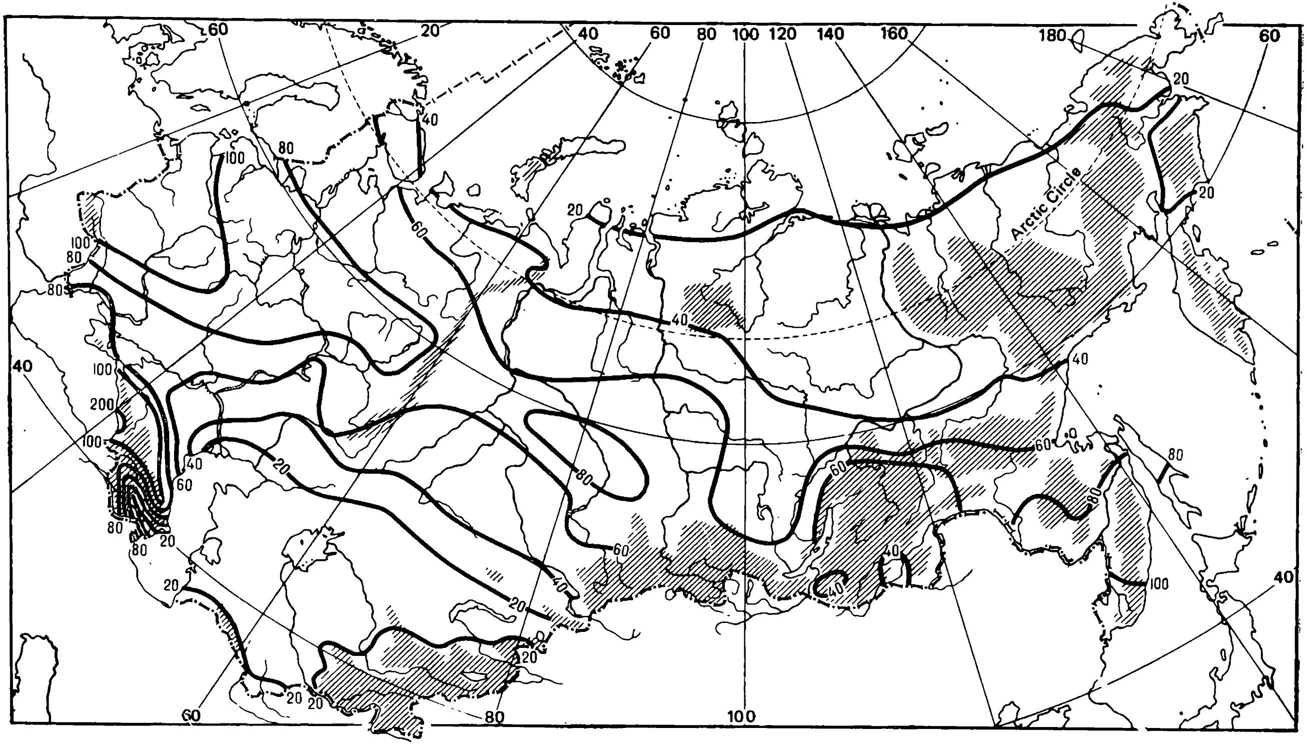

Fig. 31. Productivity of the Natural Plant Cover on the Territory of the USSR (c/ha•year)

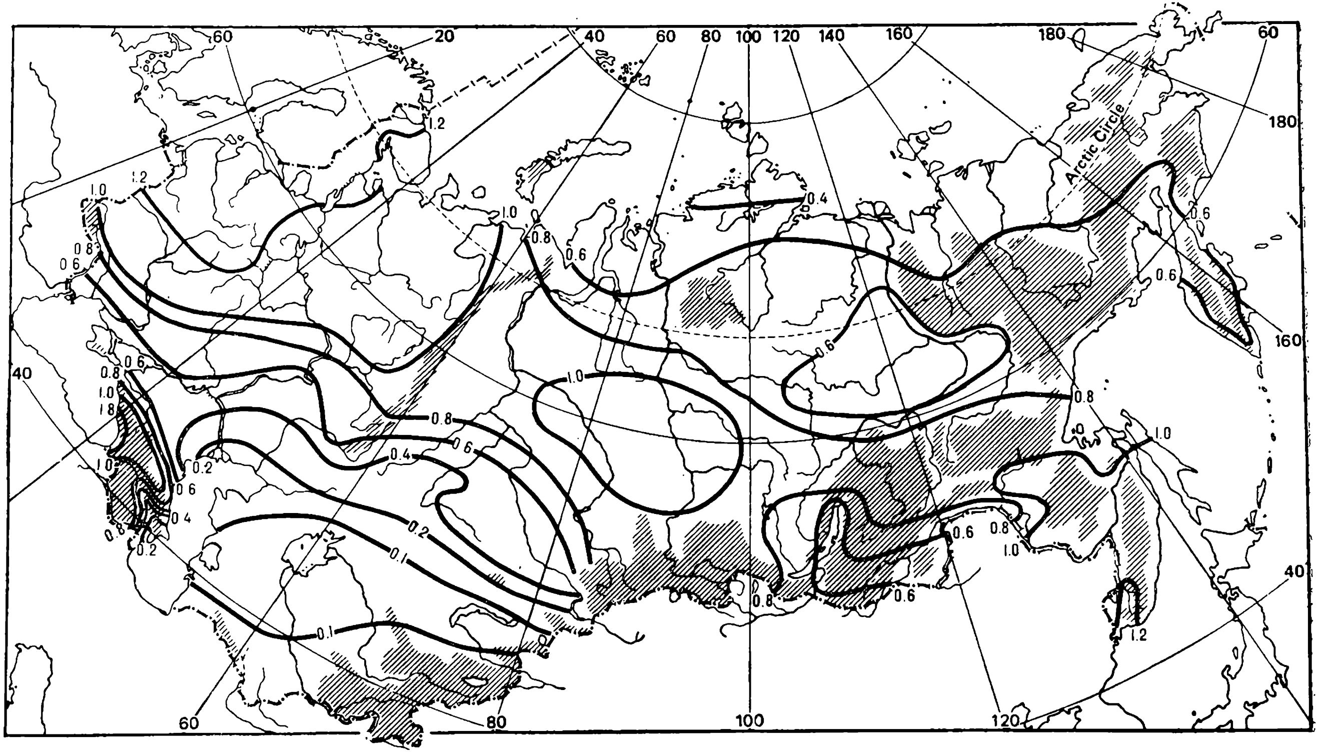

Fig. 32. Utilization of Solar Energy by Plant Covers (per cent)

ring the vegetation period. It may be seen from the map that up to two per cent of the arriving photosynthetically active radiation is absorbed in the yearly increment of organic matter. The largest coefficient of utilization of solar radiation energy (1.8-2 per cent) is found in the region of the maximal yearly increment, namely on the Black Sea coast of the Caucasus. A relatively high value of energy utilization (1.0-1.2 per cent) is found in nearly every region of the European part of the USSR, except for arid regions in the south and southeast. In the Asiatic part of the USSR approximately 1 per cent of the energy is employed in the central part of Western Siberia, in the south of the Khabarovsk Territory and in the Primorye Territory. Over the remaining parts of the country the yearly increment in biomass contains from 0.4 to 0.8 per cent of the incoming photosynthetically active radiation. In desert and semi-desert areas of Central Asia the assimilation of energy in the yearly increment of biomass ranges from 0.2 to 0.1 per cent. Let us note that in terms of their content these coefficients of utilization of solar energy are not the same as the coefficients mentioned earlier in this chapter that relate not to photosynthetically active radiation but to total radiation.

Smaller values for the coefficient of utilization of solar energy by plants in the eastern regions of our country by comparison with western ones are attributable to less favourable moisture conditions in many eastern regions, and also to a reduction in the duration of the vegetation period.

The average value of the coefficient of utilization of photosynthetically active solar radiation by the plant cover for the entire territory of the USSR is approximately 0.8 per cent. The magnitude of the corresponding coefficient for total solar radiation will be approximately one-half of that value, i.e., roughly 0.4 per cent.

In later studies Yefimova made use of a more precise relationship between the productivity of the natural plant cover and meteorological factors in calculating values of the productivity and the coefficient of utilization of photosynthetically active radiation for each continent and in compiling the corresponding maps (Yefimova, 1976). The results of her calculations are presented in Table 12.

The data in the table show that the average productivity per unit area for the five continents does not differ very much. It is evident that over a large part of the territories of each of those continents its magnitude is greatly limited by insufficient heat or moisture. The continent of South America represents an exception in this regard. Over the larger part of its territory climatic conditions are very favourable to the development of plant life.

In Australia and Africa the coefficients of utilization of photosynthetically active radiation are lower than average. This is attributable to insufficient moistening over large areas of these continents, which inhibits the utilization of solar radiation energy by plant covers.

The most favourable moisture conditions for the development of plant life are found in Europe and South America; in Europe, which is located at higher latitudes and is exposed to less solar radiation, its utilization is relatively greater.

Table 12

PRODUCTIVITY OF NATURAL PLANT COVER

Continents	Productivity		Coefficient of utilization of photosynthetically active radiation energy (as a percentage of total over the vegetation period)
	tonnes, 10^9	Centner /ha	
Europe	8.9	85	1.26
Asia	38.3	98	0.88
Africa	31.0	103	0.59
North America	18.1	82	0.94
South America	37.2	209	1.13
Australia (including islands of Oceania)	7.6	86	0.44
Land as a whole	141.1	95	0.86

There are less data on the productivity of autotrophic plants in water bodies. Data based on observations indicate that in seas and oceans the greatest volume of organic matter is produced by plankton located in the layer 30 or 40 meters deep. At greater depths the quantity of solar radiation is insufficient to make possible an active development of photosynthesis.

Productivity is substantially less in the open ocean than it is in shelf zones. It may attain maximal values in small bodies of water possessing large quantities of mineral compounds required by plants.

According to data prepared by V. G. Bogorov (1969) the overall value of productivity for the oceans is approximately 55 billion tonnes per year, i.e., approximately 15 c/ha·year. This last figure is less than one-sixth of the average productivity per unit area on continents.

On the basis of that estimate we will find that the yearly volume of productivity for the Earth as a whole is approximately 200 billion tonnes, i.e., about 40 c/ha. This corresponds to an expenditure of energy that is approximately equal to 0.15 kcal/cm^2·year, i.e., about 0.1 per cent of the volume of solar radiation reaching the Earth's surface.

One may think that these values are too small to permit the process of photosynthesis to influence the major components of the heat balance at the Earth's surface directly. In this connection it is usually believed that photosynthesis is not a substantial factor influencing climatic and weather conditions.

In recent years, however, it has been established that in our epoch climatic conditions are rather unstable, and that in this connection changes in the components of the Earth's energy balance by several tenths of one per cent of the arriving solar radiation may exert a perceptible influence on the position of the ice cover and on climates at middle and high latitudes (see Chapter III).

Accordingly, the question of the possible influence of the energy of photosynthesis on meteorological processes calls for some attention. Let us note that such an influence is limited by a relatively rapid return into the atmosphere of energy assimilated in the course of photosynthesis as a result of the general biological cycle that transforms that energy into heat.

Thus climatic conditions are influenced by differences between the energy expended on photosynthesis and the quantity of heat that is received as a result of the decomposition of organic matter. That difference may constitute only a small part of the expenditure of energy on photosynthesis and its climate-forming significance in our epoch will apparently be negligible. That influence is less clear, however, for periods during which a rapid accumulation of organic matter occurred (for example during the Carboniferous period). But even under such conditions it seems unlikely that the influence of photosynthesis on the meteorological regime was substantial.

Nevertheless, there are cases in which such an influence is quite tangible. This refers to fires that affect substantial areas] of forests. In such cases the energy of photosynthesis accumulated over long periods of time is transformed into heat so rapidly that the volume of additional inflow of heat per unit area is much larger than the energy of solar radiation. As a result, the meteorological regime on territories where large forest fires occur changes sharply, and such changes may extend to a considerable layer of the atmosphere.

Except for such special cases, however, it should be noted that the plant cover exerts its greatest influence on climate not as a result of the expenditure of solar energy on photosynthesis but in connection with the influence of transpiration on the balance of heat and of moisture (see Chapter IV).

Thus data relating to coefficients of the utilization of solar energy by plant covers are important mostly as a characteristic of climatic conditions governing the development of plant life and as indicators of the effectiveness of the photosynthetic activity of different types of plant cover.

Let us note that the data that have been mentioned concerning productivity describe the quantity of organic matter that may be utilized by heterotrophic organisms. Primary productivity (i.e., the volume of assimilation) is always substantially greater than the total productivity considered above, because a part of the organic matter is expended on plant respiration. In many cases total productivity is approximately .one-half of primary productivity.

Transformations of Organic Matter. Organic substances that are produced by autotrophic plants are nearly fully consumed by heterotrophic organisms.. A part of these organisms feeds on live organic matter while another, which largely consists of microorganisms, feeds on atrophying parts of autotrophic plants. A relatively small share of the total organic matter produced is transformed into mineral substances without the participation of living organisms. Examples of such process are provided by forest and prairie fires, in the course of which organic matter is transformed into carbon dioxide, water vapour, and certain mineral compounds.

A very small part of the organic matter produced remains in upper layers of the lithosphere and at the bottom of water bodies in the form of coal, peat, and other organic compounds.

Nearly the entire total output of autotrophic plants thus enters into various food chains.

It has been noted in Chapter I that the transmission of energy within trophic chains from one group of organisms to others is associated with large losses in energy. The ratio of the biomass of organisms to the quantity of organic matter that they consume is sometimes called the coefficient of ecological effectiveness. As a rule that coefficient does not exceed 10-20 per cent.

Small values of the coefficients of ecological effectiveness lead to a quick reduction in the mass of organisms located at subsequent links of the trophic chain. The mass of organisms located at a specific link of a trophic chain (in other words, at a specific trophic level) depends on their size.

The intensity of processes of exchange (metabolism) per unit weight of a live organism is usually the greater the smaller organism is in size. One of the causes for this principle is that exchange processes depend substantially upon the rate of diffusion of gases through the surface of organisms which increases per unit of their biomass as their size diminishes. The metabolism of a given unit of weight of microorganisms is thus many times larger than the metabolism of the macroorganisms. Metabolism also depends on the nature of physiological processes within the tissues of organisms: in the case of wood, metabolism is usually much lower than in vertebrates' tissues of a similar size.

These principles largely determine the total volume of biomass of various organisms.

The calculation of the total biomass for the Earth as a whole is a difficult problem. According to the estimates of V. A. Kovda (1969) it is close on $3 \cdot 10^{12}$ tonnes for continents. According to his data, over 95 per cent of that volume belongs to plants and less than 5 per cent to animals. Since the mass of organisms in oceans is less than the mass of organisms on land, where most of the biomass of plants is that of trees, we will find that the overall magnitude of biomass for the planet as a whole is largely determined by the distribution of forests over continents. If we consider that the total productivity of plants on continents is $140 \cdot 10^9$ tonnes we will find that the time for one cycle of organic matter on continents is approximately 20 years. It is evident that this average figure for land largely relates to forests, which constitute the major part of the biomass of plants. In other natural zones on the continents the duration of the organic matter cycle is much shorter. That cycle is shorter still in the oceans where in the case of phytoplankton it appears to be only a few days.

The overwhelming proportion of forests in the overall biomass of living organisms is associated with the fact that as autotrophic organisms trees are located at the first link of trophic chains. It is also due to the large size of each individual tree. Together with the specific properties of wood itself, this reduces substantially the rate of metabolism per unit biomass. Even though ocean phytoplankton accounts for a relatively large productivity that is comparable with that of forests, the small size of individual plankton organisms intensifies its metabolism per unit weight so considerably that the total mass of plankton is negligibly small by comparison with that of forests.

Let us assume that the total biomass of animals is approximately 10^{11} tonnes. Assuming that animals assimilate approximately 10 per cent of the total productivity of plants we will find that the average duration of one cycle of animal organic matter is several years.

The actual length of life of one generation of animals varies very widely.

The estimate of the duration of a single cycle of animal organic matter that was mentioned earlier refers to those groups that constitute the major part of biomass. Unfortunately, the nature of the distribution of biomass among different groups of animals is not clear.

Invertebrates represent the largest component of biomass of animals, and among them on land the most important ones include organisms living in the soil. The zoological mass of large animals per unit area is relatively low. Thus the calculations of J. Huxley (1962) indicate that while in African savannas the biomass of large wild animals may attain 15-25 t/km^2, this figure is 1 t/km^2 in forests at middle latitudes, 0.8 t/km^2 in tundras and 0.35 t/km^2 in semi-desert areas.

Unlike in the case of animals and plants a much more precise estimate may be obtained of the biological mass of the human population and of its energy consumption in the course of feeding.

For the current size of the human population (over 4 billion persons) its biomass is approximately $0.2 \cdot 10^9$ tonnes. Assuming that each person expends an average of $2.5 \cdot 10^3$ kcal of energy per day, we find that the total consumption of energy by the human population is approximately $1.8 \cdot 10^{15}$ kcal/year. This magnitude corresponds approximately to the current productivity of agriculture, even though it does not meet the requirements in food products of the world's population in all regions of the world.

Man occupies one of the last positions in trophic chains, and consumes both the primary output of autotrophic plants and the biomass produced by many herbivorous and carnivorous animals. Accordingly, his position within the energy pyramid of food lies in its upper floors which are much more narrow than the pyramid's base describing the productivity of autotrophic plants. From the data that has been mentioned it is easy to find that at the present time man consumes approximately 0.2 per cent of the primary production of the organic world. A few thousand years ago this figure was much lower than 0.01 per cent, and in the near future it is expected to grow substantially in connection with the rapid growth of population.

The dimensions and shape of the pyramid describing the exchange of energy in the biological cycle on our planet constantly change in accordance with the development of agricultural activities required to provide food to the world's growing population.

As man consumes a growing part of primary biological production he also expends a large and ever increasing quantity of technical energy which represents a new source of heat for our planet. The data in Chapter X show that this raises the prospect of a substantial change in climate, which for the first time in the Earth's history begins to depend substantially on the activities of organisms.

The Carbon Cycle. The creation of organic matter is based on the absorption by plants of carbon dioxide from the atmosphere and hydrosphere.

At the present time carbon dioxide is one of the relatively small components of the atmosphere. In the present age the atmosphere contains approximately $2.3 \cdot 10^{12}$ tonnes of carbon dioxide, and this represents 0.032 per cent of the atmoshpere's volume (volume percentage). Much more carbon dioxide is contained in the hydrosphere (primarily in ocean waters) in which approximately $130 \cdot 10^{12}$ tonnes of carbon dioxide is dissolved. A continuous exchange of carbon dioxide takes place between the atmosphere and the hydrosphere in the form of a molecular diffusion.

Measurements of the concentration of carbon dioxide in the atmosphere show that it varies little in different geographical regions and also at different heights within the troposphere. The relative stability in the content of carbon dioxide in the atmosphere, unlike that of water vapour, is explained by smaller changes in the capacity of sources and sinks of carbon dioxide on the Earth's surface and by the absence of a substantial dependence of carbon dioxide content on temperature. Data based on observations indicate that concentration of carbon dioxide increases slightly near the Equator and declines by approximately 0.005 per cent at high latitudes. These changes are attributable to the higher capacity of carbon dioxide to dissolve in the cold waters of high latitudes by comparison with the warm waters of the tropics. As a result of this, at high latitudes the atmosphere loses a part of its carbon dioxide, which is dissolved in oceans. From there surpluses of carbon dioxide are carried by cold deep streams to low latitudes where it is returned to the atmosphere. The magnitude of the flow of carbon dioxide between the Equator and the North Pole that is produced by this mechanism is approximately $2 \cdot 10^{10}$ t/year (B. Bolin and C. Keeling, 1963).

There is a constant exchange of carbon dioxide between the atmosphere and hydrosphere on the one hand, and between living organisms and the lithosphere on the other.

The major components of the carbon dioxide cycle are determined by the biological processes. In the course of photosynthesis carbon dioxide is absorbed by autotrophic plants whose life activities (in particular respiration) return almost completely the carbon dioxide that they have absorbed into the atmosphere and into the hydrosphere while a small quantity of carbon goes into the lithosphere.

A certain quantity of carbon dioxide enters the atmosphere from the depths of the Earth's crust as a result of volcanic eruptions, from mineral springs, etc. Carbon dioxide is expended in the course of the wind erosion of silicate rock and in connection with the formation of various carbon compounds.

Thus there are two cycles within the carbon dioxide balance in the atmosphere: a biological cycle and a geological one, and within each there is an inflow and expenditure of carbon dioxide. Available data indicate that the biological components of the yearly carbon dioxide cycle are substantially larger than the geological ones.

On the basis of the data on the productivity of plants cited earlier, we find that the expenditure of carbon dioxide on photosynthesis each year is approximately $3 \cdot 10^{11}$ tonnes per year. A similar volume of carbon dioxide is formed through the process of respiration and as a result of the decomposition of living organisms. A far smaller amount of carbon dioxide is obtained from the Earth's crust, namely around 10^8 tonnes. A similar quantity of carbon dioxide is expended in different geological processes (G. Plass, 1956; D. Müller, 1960; H. Lieth, 1963, and others).

Living organisms, the products of their life processes and the lithosphere contain large quantities of carbon obtained from the atmosphere and the hydrosphere in the form of carbon dioxide. The quantity of carbon in the lithosphere corresponds to approximately $2 \cdot 10^{17}$ tonnes of carbon dioxide, of which the main part is bound in carbonate rock (A. Vinogradov, 1972).

While these estimates relating to components of the carbon dioxide balance and to reserves of carbon in different natural media are very rough, they make it possible to draw certain conclusions concerning the rate at which the carbon dioxide cycle occurs.

The average time of renewal of carbon dioxide in the atmosphere is approximately ten years. The time during which an accumulation of carbon in the lithosphere has taken place is very long and is of the same order of magnitude as the time of existence of the biosphere. There are grounds for thinking that in the geological past the concentration of carbon dioxide in the atmosphere differed substantially from its present volume. This suggests tha the total of all forms of inflow of carbon dioxide into the atmosphere and of its expenditure was not equal to zero in many cases.

It may be concluded, however, that the difference of that total from zero each year represents a very small part of the absolute value of the major components of the carbon dioxide balance in the atmosphere. Thus if we assume that during the last million years the concentration of carbon dioxide in the atmosphere has declined by one-half this will correspond to a change in its magnitude per year by approximately 0.001 per cent of the yearly amount produced by photosynthesis.

The influence of fluctuations in the amount of carbon dioxide on the biosphere is considered in Chapter X.

VII

The Ecological Mechanism of the Evolution of Organisms

1. The Stability of Ecological Systems

A Numerical Model of the Ecological System. In order to study the influence on living organisms of large-scale changes in the environment it is necessary to consider the stability of the ecological systems to which these organisms belong. One should keep in mind that, just as various types of organisms, ecological systems have been subjected to the prolonged evolutionary process, in the course of which the less stable systems have vanished and only those ecological systems were preserved whose stability in relation to the fluctuations of external factors was sufficiently high.

The conception of the stability of ecological systems is frequently employed in contemporary studies (M. Antonovsky, 1975; V. Fyodorov, 1975). Different authors, moreover, define the conception of a system's stability in different ways. In accordance with an approach proposed earlier (Budyko, 1974) let us examine the stability of an ecological system through the analysis of its numerical model. Let us consider a highly simplified model of an ecological system that includes only four types of organisms, namely:

(a) autotrophic plants;
(b) herbivorous animals;
(c) carnivorous animals;
(d) parasites consuming the biomass of the first three groups of organisms.

Equations representing the balance of living organic matter which have been formulated by Lotka and Volterra for that system practically coincide with those representing the energy balance.

Let us represent the biomass of a particular group of organisms per unit of area that they occupy by B, their productivity (i.e., the inflow of biomass per unit time minus the expenditure on the organisms' life processes) by Π and the consumption of biomass by parasites by D. The equations representing the balance of biomass have the following form:

for autotrophic plants—

$$\frac{dB_a}{dt}=\Pi_a-b_bB_b-D_a, \tag{7.1}$$

for herbivorous animals—

$$\frac{dB_b}{dt}=\Pi_b-b_cB_c-D_b, \tag{7.2}$$

and for carnivorous animals—

$$\frac{dB_c}{dt}=\Pi_c-D_c, \tag{7.3}$$

where b_b and b_c are coefficients describing the consumption of biomass by the organisms that are being eaten by herbivorous animals and carnivores per unit mass of the corresponding type of animal. The derivatives on the left side of equations (7.1-7.3) determine the rate of change of biomass over time.

If as a first approximation we consider the coefficients b_b and b_c to be constant, we note that equations (7.1-7.3) include nine variables. Accordingly, their solution requires that six additional relations be employed. These relations may be inferred from empirical data for various ecological systems.

For a wide range of systems it is possible to represent simplified relations having the form

$$\Pi_a=\Pi_a(B_a,\ B_b) \tag{7.4}$$

$$\Pi_b=\Pi_b(B_b,\ B_a,\ B_c) \tag{7.5}$$

$$\Pi_c=\Pi_c(B_c,\ B_b) \tag{7.6}$$

$$D_a=D_a(B_a) \tag{7.7}$$

$$D_b=D_b(B_b) \tag{7.8}$$

$$D_c=D_c(B_c) \tag{7.9}$$

The nature of these relations has been considered in many empirical studies.

As the biomass B_a of the autotrophic plant cover increases its productivity Π_a first increases slowly (so long as the leaves' surface is small), and then more rapidly, while for a large density of biomass it increases slowly once again because of the decline in photosynthesis under conditions of increased shade for lower layers of leaves. Accordingly the relation $\Pi_a(B_a)$ for a specific interval of changes in biomass has the form of an S-shaped curve. For very large values of the biomass when additional expenditures on respiration are not fully compensated by photosynthesis, Π_a begins to decline. The relation $\Pi_a (B_b)$ is determined by the fact that for a given biomass of the plant cover herbivorous animals frequently increase the productivity of plants through an acceleration of the additional growth of plants when a part of that increment is channelled to the feeding of animals.

The productivity of herbivorous animals Π_b is usually associated with their biomass B_b in a manner that is qualitatively analogous to the relation $\Pi_a (B_a)$, since for a very small and also a very large number of animals per unit area the ratio of the productivity to biomass declines, largely because of the reduction of the birthrate coefficient. Evidently the productivity Π_b depends on the biomass of the plants B_a and increases as the latter increases. A similar influence is exerted on Π_b for a given value of B_b by the biomass of carnivores B_c which in the main destroy sick and old animals, thus contributing to an increase in the birthrate coefficient. The productivity of carnivores Π_c changes as their biomass B_c increases in a manner that is qualitatively similar to the relation $\Pi_b (B_b)$ as Π_c increases with increases in the biomass B_b.

The consumption of the biomass of plants and animals by parasites is usually proportional to their own biomass, even though sometimes this dependence is stronger than the one that would be described by direct proportionality, since an increase in the density of populations facilitates the dissemination of parasites from one organism to others.

By representing equations (7.4-7.9) in a quantitative form it is possible to study the patterns of the dynamics of the populations of organisms belonging to the corresponding ecological systems.

As a result of such an analysis it is possible to arrive at the following conclusions, which describe conditions that are characteristic of many ecological systems.

1. For values of the biomass of each group of organisms that are constant over time (stationary states of an ecological system) there is often more than one solution to the above equations, each of which corresponds to different possible magnitudes of biomass components of the ecological system. But not all of these solutions correspond to a stable regime of the ecological system, i.e., one in which small deviations in biomass of one of the groups of organisms from its value for a stationary state bring about a return to the same state.

2. Ecological systems do possess a definite measure of stability in relation to deviations of values of the biomass of each system's component from their values for a stationary state. That stability is defined by the interval of changes in the system's biomass components within which a return to a stationary state of the ecological system takes place. For changes in biomass that exceed such intervals a return to a stationary state does not occur and one or more components of the system are destroyed.

3. The stability of ecological systems that include all the components mentioned above is usually greater than that of systems within which certain groups of animals are missing.

Since trophic interrelations (and also non-energy ecological relations) of components of ecological systems create a certain stability, these possess certain traits of an integrated whole (a similar point of view has been expressed earlier—C. Elton, 1930). In this connection ecological systems may influence the evolution of organisms.

The Evolution of Ecological Systems. The number of organisms in each group of living organisms forming an ecological system continually changes.

Changes in the number of animals of a particular species n over time t per unit of area that they occupy are defined by the equation

$$\frac{dn}{dt} = \alpha n - \beta n, \qquad (7.10)$$

where α and β are birth and death rate coefficients. If α and β are constant, and assuming that at the initial moment of time $n=N$, we find from (7.10) that

$$n=Ne^{t(\alpha-\beta)}. \qquad (7.11)$$

For $\alpha > \beta$ the number of animals will increase, and for $\alpha < \beta$ it will decline.

If the population declines the extinction of animals will take place shortly after the magnitude n attains the critical value n_1 at which a low number of animals does not provide for their further reproduction. As the number of animals increases an ecological crisis will occur after n attains the critical value n_2, which corresponds to the number of population that cannot be maintained through the food resources of the given locality.

In such cases it is possible that animals will destroy the sources of their food, and this will lead to the destruction of the ecological system and will create a threat to the existence of other species that enter into it. In the present context the destruction of the ecological system refers to the destruction of one or several of its major components under the influence of external factors, and also to changes in relations between major system's components that lead to the same result. The groups of organisms that survive following the destruction of the former ecological system may then form, either by themselves or in combination with organisms from other systems, a new ecological system.

Taking these considerations into account let us find from (7.11) the duration of existence of a population of animals for both the first and the second case:

$$t_1=\frac{1}{\alpha-\beta}\,ln\frac{n_1}{N}; \quad t_2=\frac{1}{\alpha-\beta}\,ln\frac{n_2}{N}. \qquad (7.12)$$

These formulae may be rewritten as follows:

$$g_1=\frac{1}{\frac{\alpha}{\beta}-1}\,ln\frac{n_1}{N}; \quad g_2=\frac{1}{1-\frac{\alpha}{\beta}}\,ln\frac{N}{n_2}, \qquad (7.13)$$

where $g_1 = \beta t_1$ and $g_2 = \beta t_2$ correspond to the number of generations of animals that have existed from the initial moment of time to the moment of extinction.

It is evident that under natural conditions fluctuations in birth and death rates are unavoidable and as a result their average values over a prolonged interval of time will not coincide with each other. Certain differences in their magnitudes will also result from fluctuations in external conditions that are either favourable or unfavourable for the given species of animals.

It follows from formula (7.13) that the existence of any population may lead to its extinction relatively rapidly in the case of a very small difference between birth and death rates.

Let us keep in mind that under natural conditions the magnitudes $ln \frac{N}{n_1}$ and $ln \frac{n_2}{N}$ are usually less than 10 (that assumption corresponds to the possibility of changes in the number of animals per unit of area that they occupy from a minimal value to a maximal value that is nearly a billion times larger). Under such a condition the difference between birth and death rates of only 1 per cent leads to the extinction of the population over a period of time that does not exceed the length of life of several hundreds of generations. Since such an interval of time is much smaller than the average duration of existence of individual species determined by the rate of evolution, it is evident that under actual conditions there occurs a regulation of birth and death rates without which a prolonged existence of populations would not be possible.

The prolonged existence of populations is thus made possible by the dependence of birth and death rates noted earlier on the size of the populations, and also by the interaction of organisms entering into ecological systems.

The most important relations of this type are represented in equations (7.4-7.9). They include a reduction in the productivity of excessively numerous populations as a result of insufficient food, and an increase of the death rate of animals within these populations as a result of an intensification of the activity of parasites and predatory animals. The inverse principles influence the dynamics of low populations.

As a result of the influence of ecological factors the ability of living organisms to struggle successfully for their existence is largely determined by their adaptation to conditions that contribute to the stability of ecological systems.

It may be expected that particular species possessing a number of advantages with regard to others that are already adapted to a given ecological system may disrupt the system's equilibrium, destroy it, and consequently find itself threatened by extinction. Thus, for example, an increase in the size of predators preying on herbivorous animals and an increase in the aggressiveness of these predators may result in a destruction of the objects on which they feed and in a subsequent extinction of the predators. Similarly, the development among the herbivorous animals of effective means for defending themselves against predators may result in such a large growth in their numbers that the plant cover in their locality will be destroyed.

Generally it may be stated that the criterion determining the outcome of natural selection is not necessarily an increase in the size of the population of the given species, for such increase may in fact reduce the stability of the corresponding ecological system.

It has been noted earlier that all ecological systems exist under conditions of continuous fluctuations in the size of plant and animal populations entering into them. Such fluctuations may occur both as a result of changes in factors within the external environment and in situations in which external conditions are stable (through autooscillatory processes of change in population numbers). As a result of this ecological systems that are not very stable will be less durable, and following their destruction the populations entering into them will confront the threat of extinction. Ecological systems possessing a large measure of stability have more chances of preserving the populations that enter into them over long periods of time.

In this connection it should be noted that natural selection has a tendency to preserve organisms whose evolution increases the stability of the ecological systems. At the same time natural selection supports the existence of more stable systems and eliminates less stable systems which in this case are superorganisms, also participating in the struggle for existence.

Numerical modelling of energy linkages within ecological systems leads to the conclusion that many systems have achieved a high degree of stability at early stages of their existence, and that subsequently stimuli for their further development have come to be highly limited. Such systems, which include ancient forms of organisms, are especially numerous in the high seas, where the environment has always been relatively more constant.

The rapid development of the evolutionary process leading to major changes in living organisms appears to be less probable in relatively stable environmental conditions, that is, in the case of stable ecological systems. Since changes in the organisms that enter into them must be coordinated in a certain measure this reduces substantially the probability of a consolidation of mutations for each individual organism.

Nevertheless, even the most stable ecological systems evolve both as a result of changes in their natural environment (for example, as a result of changes in the gas composition of the atmosphere) and as a result of the operation of purely biological factors. The rate of evolution of such systems appears to be minimal.

In the presence of stable ecological conditions the evolutionary process takes the form of microevolution, i.e., it is limited primarily to the restructuring of low taxonomic groups of living organisms. Over sufficiently long intervals of time that process could bring about changes in higher taxonomic groups. In this connection one should accept the view that there does not exist a sharp boundary between microevolution and macroevolution (E. Mayr, 1963, S. Schwartz, 1969, and others). The rate of emergence of new taxonomic groups, however, is then relatively low.

Changes in the environment that disrupted the stability of ecological systems have played an important role in accelerating the evolutionary process. If these disruptions were relatively large, they could bring about the extinction of many groups of living organisms whose ecological niches sometimes remained empty for prolonged periods of time. Less severe disruptions of the stability of ecological systems also resulted in the extinction of various organisms whose ecological niches were immediately filled with other organisms. In such cases an increase in the rate of evolution occurred whose intensity reflected the intensity of changes in the environment.

Let us consider an example that illustrates the influence of changes in external conditions on the successive replacement of fauna.

Shortly after the beginning of the Tertiary period South America was separated from other continents. As a result its fauna developed over a prolonged period of time under conditions of isolation. Within that fauna an important role was played by representatives of two orders of herbivorous animals, namely litopterna and notoungulates. Towards the end of the Pliocene epoch South America was linked to North America and at that time numerous mammals penetrated into South America that differed substantially from the animals that then entered into the ecological systems existing there. This resulted in a rapid extinction of many species of South American fauna, including the overwhelming majority of species from the orders mentioned above. These species were replaced by tapirs, horses, deer, mastodonts, and other animals that formed stable ecological systems with the predators that came with them from North America, such as pumas, jaguars, and bears.

Throughout nearly the entire Tertiary period the herbivorous animals of South America entered into ecological systems that included predators from the order of mansupials that appear to have been less active than the placenta-bearing predators of North America. It is therefore natural that the predators of South America became extinct at the same time as the animals on which they were preying.

The most widely accepted explanation for the extinction of many species of endemic animals in South America towards the end of the Pliocene is that these animals were displaced by more progressive animals from North America (E. Colbert, 1958). Such an explanation of the causes of their extinction is oversimplified. A more accurate explanation refers to the destruction of ecological systems in South America after they ceased to be isolated.

By using the equation for the energy balance of ecological systems, it would be possible, in principle, to estimate the rate of extinction of different species in quantitative terms and to identify which factors produced the extinction of herbivorous animals in South America: these factors include their destruction by new species of predators, a trophic competition with the herbivorous animals that arrived from North America, contamination by parasites brought by new groups of animals, and combinations of these various causes.

Aside from disruptions of the stability of ecological systems through the spread of living organisms, the inverse process is also possible, in which the spread of living organisms is a consequence of the destruction of ecological systems. That process was studied by M. S. Gilyarov (1966).

The example of changes in the fauna of South America at the end of the Pliocene epoch that we have examined is an example of a large-scale ecological crisis that manifested itself in a relatively rapid destruction of ecological systems over a large territory. Such crises usually arise under the influence of external factors and are accompanied by the extinction (or else substantial reduction in numbers) of organisms occupying a leading role within the corresponding ecological systems.

The high stability of ecological systems made large-scale ecological crises infrequent during the pre-Holocene history of the biosphere. Nevertheless they appear to have exerted a major influence on the macroevolution of organisms in ways that will be considered in the next section of this chapter.

The Role of the Size of Organisms. Let us consider the question concerning the relevance of the size of living organisms on their evolution. It has been noted above that the basis of the energy system that links living organisms of the biosphere is the flow of energy channelled from autotrophic plants to various heterotrophic organisms. Large losses of energy at each link in trophic chains cause the flow of energy decline substantially as it is transferred from link to link. It vanishes completely at the stage of the mineralization of the organic matter of living creatures that close the food chain.

The structure of this flow of energy is characterized by a substantial heterogeneity. In a number of links in trophic chains the energy of a living substance is concentrated in organisms of a large size while in other links it is distributed among many small living creatures, and the transfer of energy may take place in both directions (for example, the feeding of some species of whales on plankton and the consumption of the biomass of whales by various microorganisms).

The size of living organisms, which determines the quantity of energy that they contain and the quantity of energy required for sustaining their life processes, is a major factor in biological evolution. It has been noted already a long time ago that those types of living organisms whose development progresses usually increase in size (Kopp's rule).

The role of size in the ecology of the plant cover of tracts of land is explained by the advantages of large plants in receiving the largest quantity of solar energy compared with small plants that often grow in shade. In the case of open plant covers in dry regions large plants with a developed root system are usually able to receive water from deeper layers within the soil, and large succulents have a large volume of water accumulated in their tissues in relation to the area in which evaporation takes place. The longevity of plants that are large in size is usually greater.

In the case of animals an increase in the size of their body adds to their stability in relation to changes in the thermal regime and reduces the danger of encounters with other animals. In most cases this greatly increases their longevity.

At the same time in both plants and animals increases in size also increase the amount of energy that is required for their life processes, and this limits the maximal possible size of living organisms.

It should be noted that for many plants and animals the trophic energy received from the environment may be approximated as being proportional to their surface, while the energy that is expended on life processes is proportional to their mass. Since in average conditions both these flows of energy are equal to each other, the intensity of the inflow of energy from the environment per unit area of the organism must increase proportionally to its size. Evidently there are limits to increases of the corresponding inflow of energy, and therefore a limit to possibilities for increasing the size of organisms.

Changes in the inflow of energy that is needed by living organisms not only limits the maximal size of organisms but also establishes a size for each organism that is optimal for the ecological niche they occupy and for which the difficulties in maintaining an energy balance are compensated by advantages resulting from increases in the size of the organism.

It should be pointed out that populations possessing identical biomasses per unit area but consisting of organisms of different sizes are characterized by different stability in relation to changes in conditions of the environment.

Let us consider that question for two populations of animals possessing an identical energy base for feeding but differing substantially from each other in size and longevity. Under such conditions the productivity of the population of the smaller animals will be greater than that of the larger animals. In particular, this is associated with the higher intensity of metabolism in the case of the first group of animals. Calculations based on equations for the dynamics of populations show that increases in death rate corresponding to loss of an equal amount of biomass for small and large animals lead to a much more rapid extinction of large animals (see Chapter VIII).

Thus large animals with a low coefficient of reproduction of biomass and possessing a number of advantages over small animals in the stable external conditions find themselves at a disadvantage when unfavourable changes take place in the environment.

One would therefore expect that for the examples of substantial changes in external conditions that have been considered many species of large animals have vanished, since they are the most vulnerable elements of ecological systems when their stability is disrupted. It is probable that this explains the filling of ecological niches of large-sized organisms with new and more progressive forms, while many small organisms continue to retain their ecological niches even following considerable changes in environmental conditions.

2. Critical Ages in the History of the Biosphere

Changes in Climate. Changes of climatic conditions have served as an important factor in the evolution of living organisms.

In the Cenozoic era, and especially in the Pleistocene epoch, changes in the thermal regime have exerted marked influence on the evolutionary process. As has been noted earlier these changes were declines in temperature in all regions of the globe that were especially pronounced at middle and high latitudes. The process of cooling became more rapid towards the end of the Tertiary period and led to a sharp reduction in temperature at the time of the ice ages of the Pleistocene epoch.

The question of influences produced by changes in climate on the evolution of vertebrate animals living on land was considered in the studies of W. Mathew (1915). As a result of an analysis of paleontological data and of data concerning the present geographical distribution of animals Mathew concluded that changes in climate have exerted a substantial influence on the evolution and distribution of animals.

Mathew believed that in the most recent geological periods new forms of life emerged primarily in the Holarctic region, from where they spread radially into all continents. He concluded that the conception that is widely held in zoogeography about the past existence of different "bridges" between continents is superfluous, and that the process of the spread of animals may be explained in terms of the present configuration of continents by considering their possible elevation and sinking within ranges corresponding to the depth of continental shelf zones.

With regard to the mechanism of changes in climate Mathew accepted Chamberlin's hypothesis that the increases and decreases in the levels of continents that occurred from time to time led to changes in climatic conditions, and that as continents became higher a cooling process developed at high latitudes that subsequently spread to low latitudes. In his opinion such changes in climate intensified the evolution of animals and contributed to their spread from the Holarctic region to the tropical zone and to the continents of the Southern Hemisphere. It should be noted that the interest in Mathew's studies is largely of an historical nature.

Among the later studies of the influence of changes in climate on the evolution of living organisms let us note those of D. Axelrod and of H. Baily (Axelrod, 1967; Axelrod and Baily, 1968). In the second of these studies Axelrod and Baily have examined the problem of causes for the extinction of dinosaurs towards the end of the Cretaceous period.

In the opinion of these authors there occurred a rise in the level of continents at that time, which produced a desiccation of many small inland seas. As a result the temperature of the air declined perceptibly at middle and high latitudes, and seasonal changes in temperature became more pronounced. Axelrod and Baily believe that the principal cause of the extinction of dinosaurs is an increase in the range of oscillations in temperature in the environment, for which the dinosaurs were not adapted by their preceding evolutionary development.

In the first of these studies Axelrod proposed a similar explanation for the known fact of the extinction of many large mammals in the course of the Pleistocene epoch. In that case he believed that the influence of cold winters at high and middle latitudes on the survival of the progeny of animals was especially important and that this led to the extinction of a number of species of mammals during ice ages.

Without mentioning other studies on the influence of changes in the climate on the evolution of living organisms let us note that the views of various authors concerning the nature of the corresponding mechanism differ substantially. For example, a diversity of hypotheses exist concerning the causes of the extinction of dinosaurs, including the presumption that the intensity of cosmic radiation increased during the age when they became extinct, that the air temperature rose above the threshold that they could bear, and that mammals destroying dinosaur eggs became more active. All these hypotheses do not have any particularly evident foundations, and as a result it is not possible to consider that this problem has been solved. Certain considerations concerning the way in which this question may be understood are noted below.

Similar difficulties exist in identifying the causes for the extinction of many large mammals during the Pleistocene epoch. It is widely known that particularly large changes in fauna occurred towards the end of the Pleistocene when the Würmian glaciation was nearing its conclusion. At that time the composition of the animal world in extratropical regions underwent a considerable change.

Together with the view that the influence of changes in climate has been decisive other hypotheses exist concerning the probable causes of the extinction of large mammals towards the end of the Quaternary period. In one of the studies of the present author (Budyko, 1967) a numerical model was employed to clarify the causes of the extinction of many large mammals at the end of the Pleistocene epoch. This model is based on a balance of the biomass of animals being pursued by hunters of the Upper Paleolithic. The calculations (they are presented in the next chapter) have shown that the hunting of large animals could have led to their extinction over a period of time that did not exceed the duration of the Upper Paleolithic.

With regard to the conception that has been mentioned above the close relation that exists between the rate of the evolutionary process and the stability of ecological systems, a major issue concerns the influence of changes in climate on ecological systems.

It is widely known that the geographical distribution of plants and animals is of a zonal nature, and that specific geographical zones include ecological systems similar in many respects. Already in the studies of Dokuchayev it was established that a close relation exists between geographical zones and climatic conditions. Subsequently that relation was studied in the works of A. A. Grigoriyev and the present author (the findings of these studies are presented in Chapter V).

Since geographical zonality is determined by climatic conditions, it is apparent that fluctuations in climate cause corresponding changes in the system of geographical zones.

It was noted in Chapter V that the boundaries of the major natural zones are determined by the total precipitation and radiation balances at the Earth's surface. Since many biological processes depend substantially on thermal conditions one should take into account the fact that the radiation balance is related to the thermal regime in a specific way. A comparison of yearly values of the radiation balance with the average air temperature during the vegetation period has confirmed the existence of a close relationship between these variables (Budyko, 1971). It may be expected that global changes in climate cause specific changes in the nature of that dependence.

Let us consider the manner in which conditions of geographical zonality change as a result of fluctuations in climate, for example, as a result of changes that took place during the Quaternary glaciations.

While the data in Chapter III have shown that the development of glaciations was attributable to specific changes in the radiation regime, calculations show that following the appearance of ice covers at low latitudes the radiation balance of the Earth's surface outside the glaciation zone changed relatively little. Since in the ages of glaciation air temperature at middle latitudes declined substantially it is evident that the relation that exists today between the radiation balance and average temperature during the vegetation period was changed during these ages.

It may be expected that as air temperature declined many geographical zones gradually shifted to lower latitudes. Shifts in zones could also take place as a result of changes in the quantity of atmospheric precipitation, which as a result of glaciation increased in some regions and declined in others.

Let us observe that changes in climate in the Quaternary period were sufficiently slow to permit shifts not only in areas of distribution of animals but also shifts of areas of the majority of plants to regions with favourable climatic conditions (in particular, this may be seen from the fact that as a result of a relatively very short change to a warmer climate in the 1920s and 1930s a perceptible change occurred in the plant cover at high latitudes) (A. Grigoriyev, 1956).

It should be kept in mind, however, that changes in geographical zones associated with glaciations were not limited to their shifts in space. Declines in air temperature in more or less constant radiation regimes produced new types of climatic conditions, and thus led to changes in the plant cover and other components of the landscape in the same geographical zones.

At the present time there are many examples of differences in landscape within the boundaries of the same geographical zones with various combinations of conditions of thermal and radiation regimes.

Such differences exist in regions in which during the warm season air temperature is substantially lower than the average values at the corresponding latitudes. In particular, on the Island of Sakhalin a low summer air temperature is combined with a large inflow of radiation energy. As a result, the plant cover on Sakhalin is characterized by a number of specific features, including the existence of gigantic forms of many plants. One might suppose that these forms emerged as a result of a combination of a relatively high productivity attributable to a large inflow of radiation and a relatively low expenditure of organic matter on respiration due to low temperatures.

This example shows that shifts of geographical zones due to changes in climate have led to specific changes in many natural processes that influence intrazonal ecological systems. In particular, new types of ecological systems appear to have emerged during the last glaciation in wide areas at middle latitudes of the Northern Hemisphere that were free from ice cover.

It is probable that together with limited changes in ecological systems attributable to shifts of geographical zones and their partial restructuring sharp changes might also have occurred in these systems when fluctuations in climate produced a full disappearance of particular isolated sectors of geographical zones. Such cases must have occurred on islands where changes in climate could replace a forest by a tundra, for example, or a tundra by a forest, and also on coastlines and in areas adjoining extensive mountain regions. Under such conditions intrazonal ecological systems could be destroyed, and this made possible the extinction of individual types of plants and animals whose areas of distribution were limited to the corresponding sectors of geographical zones.

But it would appear that the influence of relatively slow changes in climate on ecological systems has largely led to their gradual restructuring in ways that have intensified the evolution of living organisms. This conclusion is confirmed by data collected by B. Kurten (1965), who has established that the rate of evolution of mammals in the Pleistocene epoch exceeded considerably the rate of their evolution in the Tertiary period.

Short-Term Fluctuations in Climate. The evolutionary process could depend not only on slow changes in climate but also on short-term sharp fluctuations in climatic conditions. Such a possibility results from the enormous time scale of the periods marking nature's development in the geological past, which are not comparable to the scale that is accessible to direct observation by man.

Over such prolonged time intervals it was unavoidable, for example, that there were epochs characterized by an intensive volcanic activity that greatly exceeds its current intensity. Such epochs must have been accompanied by strong fluctuations in climatic conditions.

S. I. Savinov (1913), H. H. Kimball (1918), N. N. Kalitin (1920), and other authors have established that, following the explosive volcanic eruptions, increases in the mass of aerosols in the stratosphere reduce substantially the solar radiation that reaches the Earth's surface. The average amount of direct radiation may be reduced in such a situation by 10-12 per cent over periods of several months or years. An example of such a change in radiation is represented on Figure 33, which shows changes in the ratio of average monthly values of direct radiation under clear-sky conditions to their normals following the eruption of the Katmai volcano in Alaska in 1912. The curve has been constructed on the basis of observations of several actinometric stations in Europe and America, and shows that in individual months aerosols in the atmosphere reduced direct radiation by more than 20 per cent.

In a number of regions the reduction in direct radiation was even more pronounced. Thus, for example, in Pavlovsk (now Leningrad Region), which is located at a very great distance from Alaska, solar radiation over a six-month period was 35 per cent below normal. Similar changes in radiation took place following the eruption of the Krakatau

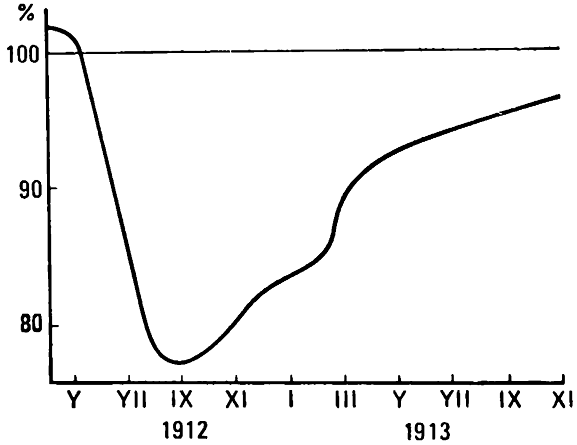

Fig. 33. Change in Direct Radiation After a Volcanic Eruption

volcano (Indonesia) in 1883. In both cases, following the eruption of the volcanos, anomalous optical phenomena in the atmosphere were observed over vast areas, which confirmed the planetary character of changes in the radiation regime resulting from the dissemination of stratospheric aerosols.

Possibilities for studying the influence of volcanic eruptions on the solar radiation regime have greatly improved as a result of the establishment of a world network of actinometric stations, mainly in the 1950s, at a time when preparations were made for carrying out the programs of the International Geophysical Year. The first large eruption of an explosive character occurred after that in March 1963 in Indonesia (the Agung volcano on the Island of Bali). A number of studies have examined the influence of that eruption on the radiation regime far more comprehensively than had been possible previously. Soon after the eruption of the Agung volcano it was found that it influenced the inflow of radiation in different regions of the globe (F. Burdecki, 1964; E. Flowers and H. Viebrock, 1965; A. Dyer and B. Hicks, 1965, 1968; M. Budyko and Z. Pivovarova, 1967). The last of these studies, in order to estimate the influence of the volcanic eruption of 1963 on the radiation regime over the territory of the USSR, processed observations of the intensity of direct radiation made at noon at a number of actinometric stations within the Soviet Union for the period from 1957 to 1966.

It was established that the average monthly values of direct radiation for 22 stations located between 40° and 68° North latitude changed relatively little between 1957 and November 1963. Following December 1963 these values declined sharply as may be seen from data presented in Table 13. A systematic change in these values beginning from the end of 1963 may be seen, together with certain fluctuations in the intensity of direct radiation from month to month, which is apparently due to the instability of atmospheric circulation.

Table 13

DEVIATIONS FROM LONG-TERM AVERAGES IN THE INTENSITY OF DIRECT RADIATION ON THE TERRITORY OF THE USSR
(per cent)

Years	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII	Over the year
1963	—1	3	—2	2	2	2	2	1	0	—1	—2	—10	0
1964	—10	—3	—6	—7	—2	—1	—3	—2	—3	—6	—8	—16	—5
1965	—15	—7	—11	—5	—4	—3	—3	—2	—5	—2	—8	—12	—6
1966	—10	—9	—7	—4	—2	—1	1	0	—2	—4	—7	—7	—4

Since the intensity of direct radiation in winter months declines much more than during summer months it follows that that decline is explained by changes in the transparency of the atmosphere rather than by fluctuations in the solar constant.

With the help of formulae of atmospheric optics it is possible to calculate from data on changes in direct radiation the changes in total radiation that exert a direct influence on the atmospheric thermal regime.

The influence of changes in radiation on the thermal regime following certain volcanic eruptions has been examined in a number of empirical studies; they have established that following major eruptions the Earth's average temperature declined by several tenths of a degree over periods of time ranging from several months to several years (W. Humphreys, 1929; J. Mitchell, 1961, 1965).

Let us examine the influence of changes in radiation on temperature after volcanic eruptions.

We shall assume that the rate of change in average air temperature at the Earth's surface is proportional to the difference in temperature at a given moment of time T and the temperature corresponding to stationary conditions T_r, i.e.,

$$\frac{dT}{dt} = -\lambda (T - T_r), \tag{7.14}$$

where λ is the coefficient of proportionality.

Denoting the temperature at the initial moment of time by T_1 we obtain from (7.14)

$$T - T_r = (T_1 - T_r)\, e^{-\lambda t}. \tag{7.15}$$

In arriving at an approximate estimate of the magnitude of λ we rely on data relating to the yearly course of solar radiation and temperature for the Northern Hemisphere, neglecting the interaction between climatic conditions in the Northern and Southern hemispheres.

Taking into consideration the fact that the ratio of the amount of solar radiation arriving at the outer boundary of the atmosphere of the Northern Hemisphere in warm seasons (April-September) to the average yearly magnitude of radiation is 1.29, we find from the ratio of changes in temperature and radiation that has been mentioned that in the absence of heat inertia such changes in radiation would produce increases in temperature of approximately 40°C. The observed difference in average temperature in the Northern Hemisphere during the warm season and over the year is 3.5° C. Considering, in accordance with this, that for $t = {}^1/_4$ year $T - T_1 = 3.5°$ C and $T_r - T_1 = 40°$ C, we find from

(7.15) that the value of λ is approximately 0.4 year^{-1}.

In the light of that estimate let us calculate the change in average temperature during the year following a volcanic eruption as a result of which the average direct radiation over a year declines by 10 per cent. In such a case the reduction in total radiation will represent 1.5 per cent and the reduction in T_r will amount to 2°C. We find from (7.15) that the magnitude that describes the change in temperature following the eruption will be several tenths of a degree. This agrees well with the magnitude of average yearly temperature anomalies following large eruptions.

In determining the influence of volcanic activities on fluctuations in temperature over more prolonged time intervals (of the order of decades) interest attaches to data provided in a study by H. Lamb (1970). It contains comparisons of temperature anomalies over different time intervals with an index characterizing the average decline in the transparency of the atmosphere following eruptions.

A very large (in absolute terms) coefficient of correlation, equal to —0.94, was found for the relation between average temperature anomalies in the Northern Hemisphere over several decades (from the Equator to 60° North latitude) for the period from 1870 to 1959 and the above index. This proves that a close relation exists between volcanic activities and changes in temperature.

For individual regions a similar relation has been found to be less close. This seems to be attributable to the substantial influence of changes in atmospheric circulation in certain regions.

In the past the intensity of volcanic activity has changed within a wide range of values. This is shown by differences in the quantity of products of volcanic eruptions contained in layers belonging to different geological ages. In the studies of A. B. Ronov (1959, etc.) it has been established that during the Phanerozoic eon the intensity of volcanic activity in different periods changed by several times.

Earlier it was shown that the influence of volcanic activity on climate is associated with increases in the content of aerosol particles in the stratosphere following eruptions of a volcanic nature and that that process leads to a reduction in air temperature at the Earth's surface. It may be expected that strong short-term fluctuations in volcanic activity could alter substantially the atmospheric thermal regime.

It was noted in a number of geological studies that possible deviations of the intensity of many natural processes from normal increase as the period of time under consideration is prolonged; for example, catastrophic earthquakes which are not very probable for short intervals of time were quite probable in the course of sufficiently long periods.

Thus from a general point of view it seems natural to expect that over the Earth's long geological history there were grandiose anomalies in the characteristics of natural processes that have not been observed over the relatively short time during which mankind has existed, and especially over the very short time during which man has been studying his natural environment.

Let us consider the possibility of applying that principle to estimates of fluctuations in volcanic activities in the past.

While the influence of individual eruptions on the Earth's temperature is relatively small, in view of the limited quantity of aerosol, entering into the atmosphere following each eruption, it is evident that the Earth's temperature will change much more substantially when many eruptions of the explosive type take place over a relatively short time interval.

The possibility of such coincidences over long periods of time, which is a consequence of general statistical regularities, increases appreciably in the presence of fluctuations in volcanic activity. Similarly, the greatest quantity of aerosol entering the atmosphere during a single volcanic eruption will increase as the time interval that we examine increases for the same general statistical reasons, as well as the greatest frequency of eruptions.

A number of difficulties are encountered in attempting to calculate the greatest possible quantity of aerosol that may enter the atmosphere during volcanic eruptions over a relatively short period of time. An initial estimate of such a quantity may be based on the data presented by Lamb (1969), who has constructed a table of eruptions of the explosive type that have occurred during the 18th-20th centuries. Lamb has included into that table an estimate of the quantity of dust entering the atmosphere following each eruption, representing it in terms of a scale on which all values are related to the quantity of dust that has been produced by the eruption of Krakatau in 1883. That quantity has been given a value of 1,000 conventional units.

An analysis of the data prepared by Lamb shows that the quantity of dust that is added to the atmosphere during each decade also fluctuates within a wide range and that the corresponding maximal values increase as the period of time becomes longer, i.e., as the number of decades being considered is increased. This relation may be expressed in terms of the following empirical formula:

$$N = \mathrm{a} \log \frac{T}{t}, \qquad (7.16)$$

where N represents the greatest amount of dust thrown into the atmosphere over an interval of time t during a period of length T and where a is a coefficient whose value is 3,000.

This formula may be employed with values of T considerably larger than t.

It follows from formula (7.16) that the greatest quantity of dust that enters the atmosphere over a decade ($t=10$) for $T=300$ is equal to 4,400, which accords well with the maximal value in the totals calculated by Lamb.

If we use that formula as a basis for extrapolations over large periods of time, we find that for $T=10^4$ $N=9{,}000$, while for $T=10^7$ $N=18{,}000$.

Thus for long periods of time it is possible that during a decade there will be a series of volcanic eruptions that are equivalent to nine and even eighteen eruptions of Krakatau in terms of the quantity of added aerosol.

One should not exaggerate the reliability of that estimate, which is based on an extrapolation procedure. At the same time one may assume that this value describes correctly the order of magnitude of a possible maximum of volcanic activity.

Let us consider how one can examine changes in the Earth's temperature in the case of many volcanic eruptions.

If in the course of a decade approximately ten eruptions comparable to Krakatau occurred, the average reduction in direct radiation during that time would be not less than 10-20 per cent. In such a case the total radiation would decline by 2-3 per cent, and the average temperature at the Earth's surface would decline by 3-5°C.

If 20 volcanic eruptions were to occur during a decade the radiation regime would change more substantially. Estimates based on formulae of atmospheric optics show that in such a case the total radiation would decline by not less than 5 per cent. Such a reduction in radiation should lead to a reduction in the average temperature at the Earth's surface by 5-10°C, and that cooling would be disseminated both to the atmosphere and to the upper layer of oceanic waters.

In spite of the approximate nature of these data it is possible to conclude that volcanic eruptions have occasionally created perceptible short-term but palpable declines in temperature that were subsequently disseminated throughout the entire globe.

It is important to note that the probability of sharp short-term declines in temperature increased substantially in ages in which mountains were actively formed, when volcanic activities increased. In such conditions reductions in temperature could apparently be larger than the estimates that have been mentioned.

In the presence of a stable climate with a relatively small difference in temperature between the poles and the Equator, which existed over the entire Mesozoic era and the greater part of Tertiary period, even such large short-term reductions in temperature could not bring about glaciation because of the high initial temperatures at polar latitudes.

Global Ecological Crises. Data produced by paleontological studies have shown that the process of extinction of large taxonomic groups of animals (orders, sub-orders and families) are characterized by the fact that groups of animals that differed in terms of their biological characteristics and that occupied altogether different ecological niches often became extinct over intervals of time that were short in a geological sense. It follows from statistical considerations that such phenomena cannot be the result of the struggle for survival in unchanging external conditions.

As an example let us consider the well-known case of the simultaneous extinction of large groups of animals that took place towards the end of the Cretaceous period. At that time five out of ten existing orders of reptiles became extinct including land dinosaurs (two orders), flying pterosaurs, and also ichthyosaurs and plesiosaurs living in water bodies. These five orders included 35 families that had very many species (E. Colbert, 1965).

Let us assume that the average length of existence of the reptile orders that became extinct during the Cretaceous period was approximately 100 million years and that the time of extinction of an order may be established with an accuracy of five million years. The probability of an accidental coincidence within five million years of the extinction of five out of ten orders, calculated on the basis of Poisson's formula, is approximately $\frac{1}{10,000}$. Such an extremely small value indicates that the coincidence in the extinction of these groups of reptiles was not accidental, but was in fact attributable to specific changes in the natural environment.

A similar calculation may be carried out to estimate the probability of extinction over a relatively short period of time of several families entering into a single order that were not closely related in terms of their way of existence. The probability of such an event for constant conditions of environment is also extremely small.

An examination of paleontological data confirms that in most cases large taxonomic divisions of the animal world became extinct not individually but in relatively numerous groups. Thus most extinct orders and sub-orders of reptiles and amphibians have vanished during three critical ages of geological history, namely at the end of the Permian, Triassic and Cretaceous periods. It follows from the calculations described earlier that such a coincidence would be extremely improbable in constant or slowly changing external conditions even if the duration of these critical ages was several million years.

Attempts to explain the disappearance of many large groups of animals in terms of a struggle for existence under constant external conditions are inconsistent for other reasons as well.

It has been noted many times in the paleontological literature that mammals have occupied ecological niches of reptiles of the Mesozoic era not as a result of competition with these reptiles but only after their extinction and after the corresponding niches were freed.

Individual examples of this type include the interesting case of the extinction of phytosaurs towards the end of the Triassic period. These reptiles disappeared suddenly, freeing an ecological niche that after a certain time was occupied by crocodiles. These developed later from the same initial forms as the phytosaurs and are very similar to them. It is difficult to explain such examples unless one does assume changes in natural conditions.

Thus in order to understand the causes of large changes in fauna in the geological past it is necessary to consider not only the influence of biological causes, but also additional factors associated with sharp changes in the environment. There is no doubt concerning the enormous influence that short-term declines in temperature exerted over various forms of organic life.

It has been noted earlier that all living organisms without exception are characterized by a specific thermal zone of existence, and that in many cases this zone is relatively narrow. A sharp decline in temperature lasting several years or decades should be expected to cause the extinction of many types of plant and animal, and in the first place of those who lived in climatic conditions close to the lower boundary of their thermal zone of existence. Whenever reductions in temperature over the entire planet permitted the preservation of a particular species at least in the warmer regions, the existence of the corresponding plant or animal following the cessation of the reduction in temperature could bring about a restoration of its existence in the former areas. If, on the other hand, the cooling was sufficiently pronounced in the warmer regions as well, then the species or group of species under consideration would vanish altogether.

As we consider specific ways in which sharp declines in temperature influence the life activities of various organisms it should be noted that this influence must have been very diverse. For each organism it is possible to establish values of declines in temperature that a) lead directly to its death; b) reduce its activity to thresholds at which it perishes in its struggle for existence; c) reduce its resistance to infectious diseases as a result of which it inevitably perishes; d) disrupt the process of reproduction.

Values of reductions in temperature that correspond to these conditions may differ, but even the lowest of these has been sufficient to bring about the extinction of the given species of animals.

Of the many cases of simultaneous extinction of large groups of animals that have occurred in the geological past particular interest attaches to the very large scale on which reptiles and amphibia became extinct towards the end of the Triassic period, and the still greater scale of extinction of reptiles towards the end of the Cretaceous period. A specific characteristic of reptiles and amphibia is the absence of thermal regulation in their bodies, and this makes these animals especially sensitive to the temperature regime.

The climate of the Mesozoic era was characterized by a relatively low differentiation in thermal zonality, a higher temperature at low latitudes by comparison with present conditions, and a much higher temperature in extratropical regions. This created favourable conditions for the existence of animals that did not possess thermal regulation (poikilothermic animals) throughout the entire globe.

Towards the end of the Paleozoic era, when thermal zonality was amplified from time to time, difficulties appear to have emerged for such animals. An interesting evidence is provided by the unusual type of adaptation of certain reptiles of the Permian period (dimetrodons, edaphosaurs). These reptiles had enormous crests on their backs which, as some paleontologists (A. Romer, 1945) believe, served to heat their bodies at low temperatures. By placing such crests in a position perpendicular to the Sun's rays these reptiles were able to assimilate additional quantities of heat that were sufficient for sustaining further physical activity.

The absence of similar forms of adaptation among Mesozoic reptiles is one of a number of indications that there was a relatively warm climate at all latitudes during that era. Accordingly, the poikilothermic animals of that time may be expected to have been particularly sensitive to substantial reductions in temperature.

The reptiles that exist at the present time are remnants of a numerous and diversified fauna of the Mesozoic era that have survived following the extinction of the overwhelming majority of both large and many small forms of that fauna. There are grounds for believing that today's reptiles possess a greater resistance to declines in temperature than the animals that have become extinct, and that their ecology may therefore provide some indication concerning the upper boundaries of resistance of reptiles in relation to cold.

Zoogeographical studies have shown that existing reptiles are largely tropical and subtropical forms, since the overwhelming majority of their species are concentrated at low latitudes. The relatively few reptiles that live outside that zone are active only during warm seasons and protect themselves in various ways against reductions in temperature during cold periods. One finds daily cycles in the activities of many reptiles, which usually cease during the coldest hours of the day.

The studies of C. Bogert (1959) and his associates concerning the ecology of reptiles have established the very great influence on their existence of the need to maintain a specific body temperature. When heat is insufficient reptiles seek to warm themselves under the Sun's rays or else on some warm segment of the Earth's surface. Certain lizards also change their colour in order to increase their absorption of solar radiation.

On the basis of both laboratory and field data one can state with confidence that reductions in average temperature within the environment of many reptiles by approximately 10°C may be expected to bring about their extinction.

In this connection considerable interest attaches to the ecology of reptiles belonging to the most ancient of the still existing orders, namely of the rhynchocephalians. The only species belonging to that order is the tuatara (Hatteria) which currently lives on several islands in New Zealand. The existence of that animal in the relatively cool climate of New Zealand is sometimes viewed as proof of the relative ability of ancient reptiles to resist low temperatures (A. Bellairs, R. Carrington, 1966). But such a view is based on an erroneous conception of New Zealand's climate. This is a pronounced maritime climate in which there are no large reductions in temperature during the year. It should therefore be viewed as one that is very favourable to poikilothermic animals.

Studies have shown that the gestation of an embryo in a tuatara egg proceeds very slowly, that it requires approximately fifteen months and that it is interrupted during the coldest period of the year. Such a prolonged development increases substantially the risk of destruction of the relatively few tuatara eggs by other animals and illustrates the difficulties that confront archaic types of reptiles in surviving because of insufficiently high temperatures. It would appear that even a relatively modest decline in temperature will make the development of tuatara eggs impossible and lead to the extinction of that species.

Turning to the possible influence of sharp declines in temperature on poikilothermic animals of the Mesozoic era it should be noted that it is difficult today to reconstruct the precise criteria that determined the sensitivity of various fossil reptiles to cold and that have brought about the extinction of some forms and the survival of others. It can only be assumed that reductions in temperature must have been especially harmful for large species, and also for oceanic species that could not flee at times of major declines in temperature to warmer shelters as today's reptiles are able to do. This presumption is consistent with the known fact that a nearly complete extinction has occurred of large land reptiles as well as of both orders of sea reptiles (ichthyosaurs and plesiosaurs) at the end of the Upper Cretaceous period.

While the ability of relatively small forms of land reptiles to find shelters increased the relative probability of their survival, it may be expected that the survival of several groups of reptiles following sharp reduction in temperature occurred only in a limited number of relatively warm regions. It is clear that sharp declines in temperature were much less dangerous for animals possessing thermal regulation, i.e., for mammals and birds. In particular, these animals survived the transition from the Cretaceous period to the Tertiary period, and subsequently occupied all ecological niches that were freed by the extinct reptiles.

Attention must be given to the extent to which the extinction of reptiles at critical moments in their history was in fact simultaneous. While considerable difficulties attach to an analysis of available paleontological data in view of the low reliability of the dating of individual deposits, many types of evidence point to the likelihood that the extinction of various types of reptiles towards the end of the Upper Cretaceous period occurred at times that were relatively close to each other, even though they did not coincide.

Such a conclusion is consistent with the view that sharp short-term reductions in temperature led to the extinction of reptiles if one keeps in mind the probability noted earlier of series of short-term reductions in temperature during periods of intense volcanic activity. In such a situation the initial reductions in temperature could destroy animals that were least adapted to cold temperatures, while the subsequent ones may have led to the extinction of other groups that were weakened by earlier declines in temperature. Because of their relatively short duration such reductions in temperature would not leave direct traces in geologic formations even though their influence on the organic world was catastrophic in the literal sense of that word.

The qualitative characteristics of the extinction process, as a result of which only land forms of reptiles (primarily of small and medium size) survived, accords well with the probable consequences of short-term reductions in temperature. One may even imagine that the extinction of all members of particular orders and the preservation of certain members of others would reflect in some measure the differences in their resistance to cold. It would seem quite natural that a direct connection should exist between the evolutionary origins of reptiles entering into various orders and their relation to the thermal regimes.

The largest changes in the composition of living organisms thus appear to have resulted from sharp changes in conditions of the environment that were produced by global ecological crises. The example that has been mentioned illustrates only one of many possible mechanisms that call for further studies.

It is possible to entertain the hypothesis that fluctuations in the conditions of the environment, and especially major changes in these conditions, greatly intensify the evolutionary process. This accords with the view of Simpson (1944) who noted that the extinction of old groups of animals often coincided with periods of mountain formation and wrote: "Among reptiles and mammals the proportion of orders that seem to have arisen during times of pronounced emergence and orogeny is greater than can reasonably be ascribed to chance.... This is probably true of birds also...".* Simpson also drew attention to the reduced rate of evolution at tropical latitudes where external conditions have always been more stable.

The work of Simpson that has been mentioned emphasizes a specific form of the evolutionary process that he has called "quantum evolution". In quantum evolution higher taxonomic groups are replaced—families, orders, and classes, and the intensity of changes in living organisms is especially great. It may be expected that the mechanism of quantum evolution is particularly important in cases when ecological systems cease to be stable and the organisms that belong to them partly perish as they are subjected to rapid changes, and partly become components of new ecological systems. Presumably quantum evolution is the most active form of the macroevolutionary process.

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* G. G. Simpson, *Tempo and Mode of Evolution*, New York, 1944, p. 113.

Ecological crises produced by changes in climate have thus exerted a definite influence on the evolutionary process. That influence, moreover, appears to have been especially great at times when sharp short-term fluctuations in climate occurred.

VIII

The Ecological Mechanism of the Evolution of Man

1. *The Origin of Man*

The Difference Between Man and Animals. The hypothesis that man originated from ape-like ancestors was expressed as early as the eighteenth century. In particular, I. Kant supported that hypothesis and in the early nineteenth century so did J. B. Lamarck.

A major role in clarifying man's origin was played by the works of Charles Darwin and Alfred Wallace, which were published towards the end of the 1850s. They contained a theory of the origin of species in terms of the natural selection of the best adapted organisms. Although these studies did not consider the problem of man's evolution (Darwin limited himself to observing that the conception that he proposed may shed light on man's origin) the possibility of applying Darwin's theory to explain man's evolution became immediately apparent to many persons.

Already in 1860 the famous discussion between Th. Huxley and S. Wilberforce took place, in which Huxley defended the hypothesis that man originated from ape-like animals as a result of a process of natural selection. The 1860s saw the publication of studies by Huxley, Haeckel, Vogt, and other authors in which they developed that point of view.

The position advocated by Wallace was somewhat different. While in his first work concerned with the origin of man (Wallace, 1864) he presented views similar to those of Huxley, a few years later he came to the conclusion that natural selection could not be the only cause for man's development

(Wallace, 1869). In his opinion natural selection could bring about only a slight improvement in man's mental facilities by comparison with those of higher apes. He noted that the development of man's brain substantially exceeded the requirements of his existence at early stages of social development. In explaining this lack of correspondence Wallace assumed that a "higher rational will" influenced the process of human development. It should be noted that the assumption that non-material factors influenced the origin of man was the only such assumption that was present in Wallace's numerous biological studies (W. George, 1964). Darwin's own view concerning the evolution of man was expressed in his study entitled *The Descent of Man and Selection in Relation to Sex*, which was published in 1871. That study contained comprehensive data presented in support of man's origin from ape-like ancestors, similar in a number of respects to contemporary higher apes. Darwin pointed to the probable place of man's origin, Africa, where one finds today the two types of apes that are closest to man, namely chimpanzees and gorillas. (A similar view was expressed by Wallace.)

In his own study Darwin mentioned only in passing Wallace's argument concerning the difficulties that attach to explaining the development of man's brain and voiced his disagreement with these considerations. Apparently these disagreements were based on a comparison, carried out by Darwin, of the mental activities of man and animals. This led him to conclude that there is no qualitative difference between the brains of animals and of man.

In the light of contemporary views on this subject such a position cannot be accepted. Nevertheless Darwin's conclusion at the end of his study deserves attention: "We must, however, acknowledge, as it seems to me, that man, with all his noble qualities ... with his godlike intellect which penetrated into the movements and constitution of the solar system—with all those exalted powers—man still bears in his bodily frame the indelible stamp of his lowly origin."*

It is clear from this sentence that in a certain measure Darwin did recognize the fundamental difference between man and the most developed forms of animal life.

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* Ch. Darwin, *The Descent of Man and Selection in Relation to Sex*, Murray, London, 1871, p. 868.

Wallace survived Darwin by three decades and witnessed many new achievements in the study of man's origin. Nevertheless he maintained his original view concerning that problem to the end of his life. Because of Wallace's authoritative position and of the fact that he greatly valued the studies of Darwin and disagreed with them only reluctantly, the differences between his view and that of Darwin concerning such a major question should have attracted attention. Yet they were quickly forgotten, probably, because Wallace's assumption concerning the influence of non-material factors on the evolutionary process did not seem acceptable.

In the middle of the nineteenth century there were very few paleontological data shedding light on the origin of man. During the past century many major discoveries were made in that area, among which the discovery of remains of Australopithecuses in South and East Africa was particularly important, as was the discovery of remains of Pithecanthropes and of closely related creatures (*Homo erectus*) in Java, China, and Africa.

Data provided by paleontological studies confirmed Darwin's and Wallace's views concerning the African origin of man. But many characteristics of Australopithecuses, which were close to those of man's early ancestors, have proven to differ from those of contemporary higher apes. Australopithecuses were largely predators and hunted a diversity of animals with the help of simple weapons. Thus the use of animal food that characterizes modern man possesses a very ancient origin.

Australopithecuses existed towards the end of the Pliocene and over a part of the Quaternary period, i.e., during a period ranging from several million years B.C. to several hundred thousand years B.C. During the Pleistocene epoch there existed for several hundred thousand years representatives of the *Homo erectus* species that in many respects occupied an intermediate position between Australopithecuses and man. Recent studies have shown that the first representatives of the genus Homo may have appeared at the end of the Pliocene.

Modern man (*Homo sapiens*) appeared approximately 40,000 years ago, and it is assumed that he originated from Neanderthal man (*Homo neanderthalensis*), who existed over a more prolonged period and became extinct at approximately the same time when modern man appeared. Since later forms of Neanderthal men differed even more from contemporary men than did earlier ones, the view that modern man originated from late Neanderthal men is sometimes questioned.

The greater part of modern man's existence corresponds to the first of the cultures that he created, namely that of the Upper Paleolithic. During that time the hunting of large animals constituted the economic basis of social life. The studies whose findings are presented below have led to the conclusion that the culture of the Upper Paleolithic ended with an ecological crisis over very wide territories that was associated with the destruction by hunters of some of the animals of the Upper Paleolithic.

It was noted earlier that the development of such crises was quite unusual in the history of the biosphere, since ecological systems are usually subject to self-regulation with regard to the numbers of organisms that enter into particular food chains, and this excludes the possibility of a rapid extinction of particular types of plants and animals. The fact that this regulation was disrupted indicates that the emergence of modern man constituted a specific boundary following which man ceased to be subject to the operation of biological regularities governing the size of animal populations.

The destruction of large animals appears to have led to a reduction in the numbers of human population in several large regions until the time when hunting and the gathering of useful plants came to be replaced by animal husbandry and agriculture, that is in the Neolithic, approximately 10,000 years ago. Thus the period during which man has produced the food that he consumes represents approximately one-fourth of the overall duration of his existence.

The time corresponding to the development of a technical civilization is even shorter, and represents a fraction of one per cent of the total duration of man's history. Similarly the time corresponding to the revolution in science and technology does not exceed several hundredths of one per cent of that duration.

While available paleontological and archeological data shed light on many stages in man's evolution, the nature of the mechanism that has governed that evolution requires further research.

One of the difficulties that confront attempts to explain that mechanism was already noted by Darwin. Having concluded that man originated from ape-like ancestors he added: "The high standard of our intellectual powers and moral disposition is the greatest difficulty which presents itself, after we have been driven to this conclusion on the origin of man."* (Darwin, 1871). In order to overcome these difficulties it is important to clarify the question formulated by Wallace more than a hundred years ago.

Natural selection usually preserves those changes in organisms that are useful for their vital activities and that are actively employed in the course of their existence. Yet one has the impression that over nearly the entire history of modern man the enormous capacities of his brain were employed in only a very small measure.

There are grounds to believe that man's central nervous system has not changed much since the Upper Paleolithic. Evidence in support of this view is provided by the drawings of animals in the caves of Western Europe, which are often viewed as the highest achievements in man's representation of animals throughout his entire history. The impression that an enormous gap exists between modern man's capacities for mental activities and the actual utilization of these capacities at early stages of cultural development is derived from observations concerning the possibility of transferring the achievements of modern civilization to representatives of the most backward tribes that live at the level of Stone Age cultures. Today, when representatives of such tribes receive the corresponding education from early years, they easily enter into modern technological civilization. One may therefore think that under earlier conditions the mental capacities of such persons were employed only to a small extent.

Attempts to explain this gap within the context of the usual conception of evolution through a natural selection of the best adapted organisms meets with difficulties. Although many animals acquire as a result of domestication certain forms of behaviour that are not characteristic of their existence in a wild state, there are no analogies for such a gap in the world of animals. The assumption that an organ as highly perfected and complex as modern man's brain could have developed as a result of preadaptation, i.e., accidentally to an extent, is difficult to accept.

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* Ch. Darwin, *Op. cit.*, p. 609.

Accordingly, in order to explain the origin of man, it is necessary to take into consideration the operation of special factors that have led to a qualitative distinction between man and all animals. The nature of these factors deserves great attention.

The Role of Changes in the Environment. According to the views expressed by a number of researchers (I. Gerasimov, 1970 and others) it may be thought that the evolution of man was influenced in specific ways by changes in climate that occurred during the Pliocene and Pleistocene epochs. It was noted in the preceding chapter that a cooling occurred at middle and high latitudes during that time, which led to substantial changes in climate on all continents. It is quite probable that these fluctuations in climate served to intensify the evolution of both man and many other living organisms.

It is difficult to estimate the rate of human evolution. According to data prepared by B. Kurten (1965) the average length of existence of European species of mammals during the Pleistocene epoch was approximately one million years. That magnitude is comparable with the length of existence of species that were close to man during the Pleistocene epoch. It would thus appear that the rate at which species representing ancestors of man replaced each other was not very high. On the other hand, such an estimate of the rate of evolution appears to be rather general.

The accepted system for classifying living organisms does not generally reflect their ecological characteristics. That system is largely based on morphological and phylogenetic principles. Yet there are many examples of organisms that are similar in an ecological sense but very distant from each other in terms of their origin. Similarly, organisms that are close to each other in terms of their origins are often related to their environment in very different ways.

In the course of early studies concerning man's origin based on the theory of natural selection one of the arguments in favour of such a conception was the assertion that higher apes are morphologically closer to man than to lower apes. But it is also possible to assert that at the ecological level the differences between modern man and Neanderthal man (which some authors view as subspecies of a single species) are in fact greater than the differences between Neanderthal man and many mammals that do not belong to the primates.

Such a view rests on modern man's enormous influence on the environment. This was already evident at the time of the Upper Paleolithic (when man has destroyed dozens of species of large animals), but has reached awesome proportions in our modern age. Even though Neanderthal man existed approximately ten times longer than has modern man, he did not disrupt the biosphere's balanced ecology and in this respect was much more similar to other mammals than to modern man.

There is a tendency today to lower the rank of the taxonomic groups dividing man and those primates that are related to him. This accentuates still further the lack of correspondence between such a classification and actual differences in the ecological properties of man and of his ancestors.

In view of the scope of that difference it must be concluded that changes in man's ecology have generally occurred much more rapidly than have ecological changes for different animals. It may be supposed that that process was associated with the development of man's central nervous system. This attribute is not taken into account sufficiently in systems of classification pertaining to man and those beings that are closely related to him. Existing data show that during a period of two million years the volume of the brain of man's ancestors increased by nearly three times (Kurten, 1971), and that at the same time its structure became more complex.

One of the causes of such a rapid development of the central nervous system could have been the critical situation in which man's ancestors found themselves in the tropics when, during the Pliocene and the Pleistocene epochs, major changes in climate occurred. These changes resulted from declines in temperature at high latitudes that produced an increase in the air temperature's average meridional gradient. This served to alter the atmospheric circulation system and shifted the high-pressure belt to lower latitudes. Since there is little atmospheric precipitation within that belt, moisture conditions changed in wide tropical regions, and this led to a replacement of tropical forests in a number of areas by savannas and semi-deserts.

There is every reason to believe that before that change in climate had occurred man's ancestors lived in forests, and that their way of life differed little from that of higher primates, except for a more developed ability to walk in an erect position. It is quite likely that like all primates they were largely plant-eating animals.

The immediate ancestors of the Australopithecuses (which may be called pre-Australopithecuses) were morphologically similar to Australopithecuses and were relatively small in size. The erect pre-Australopithecuses walked much more slowly than most four-legged animals of the same size, and they also lacked such organs as large fangs and claws for attacking others and defending themselves.

In open areas where it was impossible to seek refuge in trees, pre-Australopithecuses were nearly defenceless even in relation to relatively small representatives of a large number of predators of African savannas. At the same time the extinction of many edible plants on which they had fed in tropical forests made their reliance on plant foods much more difficult.

Their low birth rate, which is characteristic of higher primates, and the long period of growth of their progeny to adulthood represented a major obstacle in maintaining their population. This made them particularly vulnerable to predators.

The principal factors that could provide for a survival of pre-Australopithecuses were their relatively developed brain and their ability to walk in an erect position which fully freed their hands.

It may be assumed that pre-Australopithecuses found themselves in extremely harsh conditions of natural selection, and that this led to a high rate of evolution in conditions of declining population. Such an evolutionary process, which G. Simpson has called quantum evolution, does not usually leave data in paleontological records because of the small numbers of the corresponding animals (Simpson, 1944).

Paleontological studies have shown that Australopithecuses were able to use the simplest instruments in defending themselves and in attacking others. By using these instruments Australopithecuses were able to attack successfully such large apes as baboons and possibly various hoofed animals. As a result, they were able to produce substantial volumes of animal food.

A number of remains of Australopithecuses indicated that their numbers were not very small, and that accordingly they were able to overcome the crisis situation that has been described above.

One should note that even in using weapons such as stones, sticks, and bones of hoofed animals groups of relatively small-sized Australopithecuses found themselves in difficult situations when they were attacked by large predators. It was equally difficult for slow moving Australopithecuses to pursue any rapidly moving animal. Their success in struggling for survival could be achieved only through their substantial superiority in intellectual capacities by comparison with all animals that attacked them or that they hunted. Accordingly, in the course of their evolution natural selection greatly favoured the development of their brain, and this, in turn, contributed to their gradually increasing success in defending themselves from predators and in hunting various animals.

The increasing mental capacities of Australopithecuses encouraged their utilization of various objects as weapons and made possible a transition to the construction of instruments of labour. That event was particularly important. In this connection a major role was played by the need of Australopithecuses for collective activities in defending themselves and in attacking other animals. It is on the basis of collective activities that elements of social organization emerged, and this led to the development of ethical conceptions determining the behaviour of individuals within collectives. The role of the labour process and of the development of language as well as of social relations in man's evolution has been described in Engels' well-known study "The Origin of the Family, Private Property and the State" (K. Marx and F. Engels, *Selected Works* in three volumes, Vol. 3, Moscow, 1977). Engels has also drawn attention to the important role of the shift in man's ancestors to animal food, which greatly intensified man's physical development.

Paleontological data provide evidence that some Australopithecuses developed in the direction of achieving gigantic dimensions. This served to greatly reduce the pressure of natural selection on the evolution of the nervous system, and led to a loss of habits bearing on the use of weapons. The corresponding lines of development led to evolutionary dead ends and to the extinction of their representatives.

At the same time there existed lines of development that led to a progressive development of their central nervous system. It may be presumed that this development was uneven, and that it intensified during ages in which the pressures of natural selection on population of man's ancestors were greater. It is quite probable that such periods coincided with the environmental changes that occurred in connection with Quaternary glaciations.

As glaciers advanced and receded, large fluctuations in climate occurred over vast areas. These fluctuations could greatly alter the conditions within which man's ancestors existed, and this confronted them with new tasks and intensified their evolution. It would appear that the last major stage in man's development took place during the Würmian glaciation, when *Homo sapiens* developed.

While these conceptions shed light on the role of changes in climate in man's origin, they are nevertheless insufficient to explain the development of man's human brain above the requirements corresponding to the needs of primitive societies. In our opinion this calls for an examination of the lack of correspondence between the technical means employed by man's ancestors and the tasks that they confronted.

Until recently the hunting of large animals and defence against predators were accompanied by substantial difficulties and risks even for representatives of civilized people possessing firearms. These difficulties were far greater for man's ancestors who relied on primitive weapons for hunting, and especially for his distant ancestors who were apparently much smaller in size and weaker.

Encounters with large predators were nevertheless inevitable. In addition, dangerous situations were encountered in the course of searching for animal food in situations when only large animals could provide a sufficient quantity of meat for more or less numerous tribes.

There is no doubt that usually man's ancestors were victorious in such encounters. In particular, this is confirmed by the findings of remains of such very large animals as elephants, mammoths, and large bears at locations where ancient men had established camps. Because man's ancestors possessed only primitive weapons they could fight large animals successfully only by developing special habits that far exceeded those that are needed by a well-equipped hunter today.

We will state a paradoxical assertion: the more imperfect were the hunting weapons available to man's ancestors the greater was the extent to which they had to be supplemented by a higher level of mental activity making it possible to overcome large animals. This is why requirements for man's mental development were very intense in the past, and this led to a consolidation of progressive developments in man's brain through natural selection.

Such a conception makes it possible to assume that man's evolution was more intensive in regions where hunting for large animals constituted the primary source of food. In particular, such conditions existed in large areas devoid of forests adjoining glaciers during glaciation periods. It is possible that it is in these regions, which are located at middle latitudes, that modern man developed.

Modern man could appear in ecological conditions whose requirements in relation to his intellectual capacities were very high, both in struggling against large and dangerous animals and possible competing against closely related primates.

Following the appearance of the *Homo sapiens* species, further progress in man's material culture proceeded more rapidly in regions in which natural conditions did not provide edible plants and called for the continued hunting of large animals.

Representatives of that species, which spread over many continents, occasionally found themselves in zones in which food could be provided through easily accessible edible plants (in humid tropics and subtropical areas) or else from animals that were unable to offer serious resistance. This last situation may have taken place at the time of the settlement of Australia by modern man, where marsupial animals represented an easy prey for experienced hunters of the Upper Paleolithic. It is possible that something similar occurred at the time of the settlement of America, whose various animals never faced major encounters with man and did not possess instincts favouring their protection against hunters (P. Martin, 1973). Cultural development was less intense in such conditions, and perhaps occasionally led to the loss of earlier cultural achievements.

These considerations suggest that as there developed a material culture that facilitated man's struggle for existence his enormous mental reserves became free. In the presence of stimuli produced by complex conditions of the external environment these could be employed for a further development in material culture. In this way cultural development became more intense, and at a particular stage this made possible the appearance of elements of the first civilizations. The development of social relations in human societies played a leading role in that process.

This progress was highly uneven in both space and time. In cases in which environmental conditions placed large stresses on human society this resulted in a high level of utilization of man's potential mental capacities. In the presence of the relatively favourable conditions of the environment that were encountered much more frequently, that level was substantially reduced.

These considerations appear to contribute to an overcoming of the difficulties noted by Darwin and Wallace in explaining man's origin.

2. The Ecological Crisis of the Upper Paleolithic

Until recently the view prevailed that man's influence on his environment began to assume large proportions only in recent decades, when technological development became more intense and our planet's population began to increase rapidly.

Recent studies have shown that already thousands of years ago man's spontaneous activities produced large changes in his natural environment that in some cases threatened the further existence of human society. The first case of this type appears to be associated with the development of the culture of the Upper Paleolithic in Europe, Asia, and America.

The Upper Paleolithic was the first culture created by modern man. Its economic basis was provided by hunting large animals with the help of weapons that made it possible to kill animals such as mammoths and the wool-bearing rhinoceros. The culture of the Upper Paleolithic lasted a very long time, at least two-thirds of the entire period of modern man's existence.

It was already noted in Chapter VII that it was during the Upper Paleolithic, which coincided with the last Würmian glaciation, that many large animals inhabiting middle latitudes of the Northern Hemisphere became extinct. Thus in Europe for example many large herbivorous animals that had been numerous disappeared, such as the mammoth, the wool-bearing rhinoceros, the steppe bison and the giant deer, as well as certain large predators, including the cave lion, the cave bear, and others. Major changes also occurred in the geographical distribution of remaining animals. For example, the large herds of northern deer that had inhabited Western and Central Europe moved to other areas. Similar changes occurred in the animal world of Asia and North America and, perhaps, of other continents.

Already in the nineteenth century individual authors expressed the view that primitive man was able to destroy mammoths and certain other animals. In recent literature this point of view has been shared by I. G. Pidoplichko (1963, etc.). Abundant data supporting the view that man played a major role in destroying the animals of the end of the Pleistocene are contained in the works of P. Martin (1958, 1966, 1967, etc.). In his view primitive man played a decisive role in the disappearance of many animals not only in moderate but also in tropical climates.

It should nevertheless be noted that many researchers are skeptical concerning the possibility of a substantial influence on the part of man's activities at such an early time on the disappearance of large and widely distributed animals.

In a monograph by K. Butzer (1964) in which this problem is surveyed it is noted that the animals that have vanished were highly specialized and in this connection could not adapt to changes in natural conditions in the late Würmian period. Butzer also notes that not only large herbivorous animals but also certain predators vanished towards the end of the Pleistocene epoch that were not objects of the large-scale hunting activities of primitive men. But he also notes that not all hoofed animals in Europe became extinct at that time, and that until recently many herds of large herbivorous animals have continued to exist in tropical countries. They continued to live and multiply in spite of intensive activities by the local population. In the view of Butzer these considerations contradict the view that it is because of man's hunting activities that a number of animal species became extinct towards the end of the Upper Paleolithic. At the same time Butzer recognizes that evidence concerning the mass destruction of large animals at many places of human settlement during the Upper Paleolithic points to the possibility that man, too, exerted a definite influence on the overall biomass balance of animals that have now become extinct.

In considering the reasons why the mammoths became extinct E. Colbert writes:

"Why did the mammoths become extinct? This question, like so many of the questions having to do with problems of extinction, is extremely difficult to answer. In fact it is probable that we shall never know the real reason for the disappearance of the mammoths a few thousand years ago, after their successful reign through Pleistocene time. Very likely the extinction of the mammoths was the result of complex causes. Man may have had something to do with it, but we can hardly believe that primitive man was of prime importance in bringing an end to these numerous, widely spread, and gigantic mammals" (Colbert, 1958).

It should be noted that in consideridng the causes of the substantial changes in animal life in Europe towards the end of the Upper Paleolithic existing studies have limited themselves to general considerations that are difficult to either prove or disprove without quantitative interpretations of the problem. For example, it may be pointed out that the high degree of specialization of animals that became extinct towards the end of the Upper Paleolithic did not in fact prevent their survival under the sharp changes in the environment that took place repeatedly during the Pleistocene.

The disappearance of a number of herbivorous animals at the same time that large predatory animals became extinct does not at all prove that man's activity did not influence changes in the animal world; the disappearance of these predators may be explained by disruptions in their food chains in connection with the extinction of corresponding herbivorous animals either as a result of natural causes or as a result of man's activity. Similarly, the argument that the continued existence to this day of herds of large herbivorous animals in tropical areas precludes the possibility of a destruction of similar animals by hunters in the Upper Paleolithic at middle latitudes is not persuasive.

Let us attempt to examine the possible influence of man's activities on the extinction of animals at the end of the Pleistocene on a quantitative basis, with due consideration of major factors that have influenced the relevant animal populations. Should such calculations confirm that the indicated activities played a major role in relation to the extinction of such animals, then the significance of other factors, including climatic changes, may be viewed as secondary.

Approximately 30,000-40,000 years ago modern man first appeared in Europe, where he replaced the Neanderthal men that existed there earlier, and on the basis of more developed equipment for producing weapons made of stone and bone created effective systems for the mass hunting of large herbivorous animals. During the Upper Paleolithic there occurred a substantial growth in Europe's population, material culture increased substantially, and the first major achievements were recorded in the field of art (for example the famous paintings of animals in the caves of France and other European countries). The Upper Paleolithic came to an end 10,000-13,000 years ago, not long before the complete termination of the Würmian glaciation.

The end of the Upper Paleolithic was associated with major changes in the way of life of Europe's population. This was marked by a dissemination of the culture of the Middle Stone Age—the Mesolithic. At that time together with the appearance of a new technique for producing stone weapons such achievements of the culture of the Upper Paleolithic as cave paintings partly came to be forgotten.

There are grounds for thinking that the end of the Paleolithic was marked by a sharp change in mankind's early history. That break appears to have been associated with the cessation of mass hunting activities of large herbivorous animals. Man was deprived of the possibility of supplying himself with food and with many needed materials and had to search for new sources of livelihood.

We will describe a calculation procedure that makes it possible to clarify with the help of data from certain paleogeographic and biological studies the manner in which hunting exerted an influence during the Upper Paleolithic on the number of animals being hunted (Budyko, 1967). That calculation procedure is based on an equation that describes changes in the population of a particular species of animals over time:

$$\frac{dn}{dt} = \alpha n - \beta n, \qquad (8.1)$$

where n represents the population of animals per unit area; dn/dt is the rate of change in population over time; α is the relative increment in population resulting from births; and β is the relative decline in population produced by the number of deaths per unit time.

Parameters α and β may depend on the magnitude n; such relationships differ for different animal species.

Calculations based on equation (8.1) as well as data from field studies have shown that the population of individual animal species in natural conditions usually fluctuates within relatively large boundaries. The periods of such fluctuations are commensurable with the length of life of one generation of animals and are usually explained by changes in feeding conditions that are produced by various meteorological conditions in particular years, epidemics, and complex interactions between the number of plant-eating and carnivorous animals that in some cases have, according to Volterra, the character of autofluctuations (S. Severtsev, 1941; D. Laek, 1954).

At the same time when considered over a period of time that is substantially larger than the average life duration of an animal, the average population of a particular species of animals should be more or less constant when on the average external conditions are constant. Changes in that average population may occur either as a result of changes in natural conditions or else in connection with the evolutionary development of the animal species being studied, or else of other species with which it interacts (Severtsev, 1941; I. Schmalgauzen, 1946).

The principles that govern changes in animal populations change substantially when the animals are being systematically hunted. In such cases equation (8.1) may be represented as

$$\frac{dn}{dt}=\alpha n-\beta_1 n-g, \qquad (8.2)$$

where g is the number of animals destroyed over a given period of time by hunters.

It is evident that for animals being pursued by hunters during the Paleolithic the coefficient β_1 characterizing mortality from natural causes was less, other conditions being equal, than the coefficient β, since these hunters were more successful in destroying sick and weakened animals than healthy and strong ones.

Since products of hunting activities represented the principal source of human food during the Upper Paleolithic in

Europe it should be assumed that $g = \gamma m$, where m is the number of persons per unit area, and γ is a relative quantity of biomass of the animal being pursued that is expended on one person over a given period of time (the ratio of the weight of the expended biomass to the average weight of one animal).

In this connection equation (8.2) acquires the form

$$\frac{dn}{dt} = \alpha n - \beta_1 n - \gamma m. \qquad (8.3)$$

It should be noted that in calculating losses of animals produced by systematic hunting activities the "principle of encounters" proposed by Volterra may not be used in studying mutual relations between the number of carnivorous and plant-eating animals.

According to that principle the number of animals killed is viewed as being proportional to their numbers. This is a natural assumption in the case of small animals that may hide from hunters pursuing them. It may be assumed that man, having mastered mass hunting methods, could always find the animals being pursued within the areas that he himself inhabited, particularly if one considers such large animals as mammoths that lived in vast open tundras and steppes. In this connection the value of γm in equation (8.3) must be regarded as being independent in relation to n until the animals being hunted survive on a territory inhabited by a particular tribe in numbers that meet the tribe's current needs.

In calculating the dynamics of populations of large herbivorous animals during the Upper Paleolithic it is necessary to estimate changes in population at that time. It is apparent that such changes are determined by an equation that is similar to equation (8.1), i.e.,

$$\frac{dm}{dt} = am - bm, \qquad (8.4)$$

where a and b are the corresponding birth and death rates.

While these coefficients depend on a number of factors, in the absence of corresponding data we must limit ourselves to a rough estimate of the average magnitude of the difference between these two coefficients: $c = a - b$, assuming that it is independent of time and of population density.

In such a case we obtain the following equation from equation (8.4)

$$m = m_0 e^{ct}, \tag{8.5}$$

where m represents the population at moment of time t and m_0 is the population at the initial moment of time.

Data based on available ethnographic studies relating to tribes that had reached various levels of historical development provide some information concerning population density in Europe at the beginning and end of the Upper Paleolithic.

With the help of such an approach a number of authors were able to conclude that population density in Europe towards the end of the Upper Paleolithic ranged from 5 to 50 persons per 100 km². These magnitudes should be compared with the much lower density of populations who had not mastered the methods of mass hunting, and for which the average figures of one person per 100 km² are generally accepted (see R. Braidwood and C. Reed, 1957; K. Butzer, 1964). Assuming that the corresponding increase in population occurred throughout the Upper Paleolithic, which lasted approximately 25,000 years, we find from equation (8.5) the value of c ranging from $0.64 \cdot 10^{-4}$ to $1.56 \cdot 10^{-4}$ year^{-1}.

While such coefficients of population growth appear to be very low today, they are characteristic of early periods in human history. These magnitudes agree well with the low magnitudes of expected longevity that have been established on the basis of studies of the skeletal remains from the Upper Paleolithic.

Let us now illustrate a calculation of changes in population of some species of large herbivorous animals inhabiting Europe's tundra in the period of the Upper Paleolithic.

According to available estimates (see Chapter VI) the food resources of tundras were sufficient to sustain a population of animals corresponding to approximately 800 kg of biomass per 1 km². If in the given area the particular species prevailed over other herbivorous animals it could utilize the major part of such food resources. The figure of 800 kg of biomass per 1 km² determined the upper limit of the possible population of the corresponding animals. One would think that for such animals as mammoths, whose population did not depend much on the predators, that limit was in fact attained in favourable conditions.

Let us examine how the numbers of mammoths should change following their subjection to systematic hunting activities. This requires an application of the following equation which is derived from (8.3) and (8.5).

$$\frac{dn}{dt} = \alpha n - \beta_1 n - \gamma m_0 e^{ct}. \tag{8.6}$$

By multiplying all terms in the equation by the average weight of an animal we shall have

$$\frac{dN}{dt} = \alpha N - \beta_1 N - \Gamma m_0 e^{ct}, \tag{8.7}$$

where N is the biomass of the animals being studied per unit of area that they occupy; dN/dt is the rate of change of that biomass over time; and Γ is the biomass of animals being hunted consumed by each person per year.

That equation must be solved for an initial condition $N=N_0$, when $t=0$. This corresponds to a situation in which the biomass of animals at the beginning of the Upper Paleolithic is equal to the limit determined by food conditions.

In order to simplify calculations that are complicated because of the difficulty of estimating the coefficient β_1, we will assume that for $\Gamma m_0 e^{ct} < \alpha N$ the right-hand term of equation (8.7) is equal to zero, i.e., the average number of animals does not change when losses from hunting are less than the increase in biomass. Subsequently, as the term $\Gamma m_0 e^{ct}$is increased to values that are equal to or greater than αN_0, we will neglect the term $\beta_1 N$ and will employ equation (8.7) in the form

$$\frac{dN}{dt} = \alpha N - \Gamma m_0 e^{ct}. \tag{8.8}$$

It is evident that both of these simplifications will lead to a certain reduction in the rate of decline in numbers of animals by comparison with reality.

Denoting by τ the duration of the period of time during which hunters do not greatly influence the population of the animals being hunted, we will be able to determine that magnitude from the relation

$$\Gamma m_0 e^{c\tau} = \alpha N_0, \tag{8.9}$$

and we will find that

$$\tau=\frac{1}{c}\,ln\frac{\alpha N_0}{\Gamma m_0}. \tag{8.10}$$

In order to determine the duration of a period of time during which the number of mammoths declines until their full extinction, it is necessary to solve equation (8.8).

Assuming that $N=N_0$, with $t=\tau$, we obtain through such a solution the dependence of the biomass of the animals being studied on time

$$N=\frac{\Gamma m_0}{\alpha-c}\,e^{ct}+e^{c(t-\tau)}\left[N_0-\frac{\Gamma m_0}{\alpha-c}\,e^{c\tau}\right]. \tag{8.11}$$

Through that formula it is possible to find the period of time that is required before N declines to zero, i.e., the given species of animals becomes extinct.

In order to carry out according to these formulae calculations of changes in the population of mammoths during the Upper Paleolithic, aside from parameters whose values have already been mentioned, it is necessary to know the magnitudes of the coefficients α and Γ. The first of these describes the ratio of the number of mammoths born in a given year to the overall population of the herd. On the basis of data relating to contemporary elephants the magnitude of that coefficient is 0.05.

The coefficient Γ may be estimated on the basis of the data derived from ethnographic observations for northern tribes whose food is largely derived by hunting for large animals. Noting that the magnitude of Γ describes, aside from the direct consumption of biomass as food, its use for non-productive losses within the economy, the value of Γ should be estimated as approximately 500-1,000 kg/year per person. In a study by S. N. Bibikov (1969), who supports the major conclusions that follow from this type of calculation, the view is expressed that such an estimate of biomass expenditure is somewhat too large. While we agree that the actual consumption of meat and animal fats was appreciably less than that estimate we nevertheless consider it to be reasonable in view of the need to expend a large volume of biomass for economic activities (fat, hides, animal bones) and of large losses associated with difficulties of a conservation.

Assuming that the average value $\Gamma=750$ kg/year and that in accordance with the estimates that have been mentioned $N_0=800$ kg/km^2, and $m_0=0.01$ km^{-2} we have applied the above-mentioned equations to calculate the characteristics of changes in the population of mammoths for two values of the coefficient c, namely $0.64 \cdot 10^{-4}$ and $1.56 \cdot 10^{-4}$ year^{-1}.

Calculations of the magnitude of τ according to formula (8.10) show that the population of mammoths declined only modestly over a time period ranging from 10,000 to 25,000 years depending on the assumed value of c, following the initiation of corresponding hunting activities. Subsequently, as may be seen from calculations based on formula (8.11), the population of mammoths began to decline rapidly and in the course of several centuries following the termination of a period of stable population they became completely extinct.

In evaluating the results of these calculations one should keep in mind that their precision is limited by the approximate nature of the values of parameters that were employed. At the same time the analysis of the equations that were employed indicate that appreciable changes in the magnitudes of calculation parameters do not alter general regularities governing the process being examined and that the major conclusion concerning the possibility of an extinction of large animals such as mammoths over a period of time comparable to the duration of the Upper Paleolithic retains its validity.

It may be easily shown that this conclusion will not be altered if we replace the dependence of the population on time that has been employed with a simpler linear relationship.

It may be noted that this conclusion will also be confirmed in cases in which the species of animals under consideration (for example, mammoths) represent only a part of the herbivorous animals feeding on a particular source of food. This merely requires that it be assumed that the ratio of biomass of the given species to the overall biomass of the animals being hunted approximates the similar ratio for the biomass of animals being killed by hunters.

Let us also note that such a conclusion may also be reached without any complex calculations, through a simple comparison of available information concerning the balance of the biomass of large herbivorous animals in Europe during the Upper Paleolithic. In such a context particular importance attaches to comparisons of the yearly increments in that biomass with the numbers of the population occupying the corresponding territory.

If, as in the case of mammoths, the corresponding increment did not exceed 4,000 kg per 100 km^2, then it is apparent that, even if it were fully utilized by hunting tribes, that increment would be sufficient to sustain a human population of not more than several persons per 100 km^2. Yet it is also clear that having achieved such a limiting value population in fact kept on growing, for it then continued for some time to be supplied with sufficient quantities of food not only through increments in biomass but also through the destruction of a part of the basic herd of the animals being hunted. As a result, the increments in biomass gradually declined and the herd population also declined until the animals were fully destroyed. This accords with the results of the above calculation.

It is possible that after mammoths had been destroyed by man over a large part of the territory that he occupied their last herds were preserved in regions of northern Asia with rare populations. In regions possessing such a harsh climate, however, large herbivorous animals could find adequate food resources only during relatively warm periods. Changes in climate in the direction of a declining temperature may be expected to have led to the extinction of mammoths in these regions.

An application of these equations may explain why primitive man was able to destroy mammoths but was unable to destroy elephants, which were still numerous until recently in a number of tropical regions.

It may be seen from formula (8.10) that the period of time during which animal species being hunted continued to exist depends on the possible reserve of their biomass N_0 and on the relative magnitude of the consumption of that biomass by hunting tribes Γ. Available data indicate that the magnitude of the biomass of large herbivorous animals in tropical savannas exceeds the corresponding magnitudes in tundras by at least ten times. At the same time the food of primitive man in tropical latitudes could include a much greater proportion of plant products than that of Paleolithic hunters in Europe, when glaciation limited sources of food to products of hunting activities.

Assuming that the magnitude Γ for hunters in tropical regions was not more than one-half of the one that we have accepted for conditions relating to the Upper Paleolithic in Europe, and employing a value for N_0 that is characteristic of savannas, it may be found from formula (8.10) that the magnitude τ for tropical areas must be approximately three times larger than the one found for Europe during the Paleolithic. Thus the population of elephants in the presence of constant hunting activities could remain relatively high for several more tens of thousands of years following the full destruction of mammoths. This means that until recently elephants did not confront a threat of extinction.

This theory makes it possible to explain why during the Upper Paleolithic in Europe only the largest of herbivorous animals became extinct. The magnitude for the increment in biomass of an animal species was primarily determined by the coefficient of birth α because for any of these species the upper boundary for their population was limited by food resources of the given geographical zone. The magnitude of that coefficient is very small for the largest animals and increases as the size of the animal declines. It was therefore those types of large herbivorous animals whose increments in biomass per unit area was smallest that were least protected from hunters of the Upper Paleolithic. The reason for the biological catastrophe that these species encountered was very simple.

Increases in the size of large herbivorous animals that occurred in the course of evolution freed them in varying measures from attacks by predators, and as this important factor in the reduction of the species population declined an appreciable fall in birthrate became possible and purposeful. When man during the Upper Paleolithic initiated hunting activities for these animals his role was similar to that of wolves and other predators in relation to smaller herbivores. But while small-sized herbivores could replace losses resulting from pursuit by predators through a high birthrate, large animals did not have such a possibility, and their extinction as a result of forms of pursuit to which they were not adapted in terms of their initial development was a matter of time.

It should be noted that such an explanation of the causes of the more rapid extinction of large animals towards the end of the Pleistocene epoch differs from the one proposed by G. Simpson (1944), who believed that it was more difficult for large animals with a less frequent replacement of generations to adapt to changing natural conditions of that time.

It is possible that in individual regions of Europe the actual period of time from the initiation of mass hunting activities to the extinction of large herbivores was much shorter than the value that may be obtained from calculations based on the above system of reasoning. This may be due to the assumptions that were made, and also to the fact that the calculations did not consider fluctuations in the number of animals resulting from diseases and especially as a result of changes in the food base, which could be quite significant in the conditions of the late Würmian period.

The existence of such fluctuations could intensify the decline of particular large herbivores in specific areas over many thousands of years, while the replacement of the species through migration from other regions was made difficult by the continuing pursuit of large herbivores by hunters of the Upper Paleolithic.

From such a point of view intensive hunting conditions should have resulted in an extinction of the largest animals with the lowest birthrate. Their extinction could contribute to a temporary increase in the numbers of smaller-sized herbivores, whose possibilities in relation to the utilization of the food base were improved. However, further increases in the intensity of hunting activities could also lead to the extinction of herbivores with larger birthrates, as was the case with northern deer in a number of regions of Western Europe. It should be added that for such animals losses deriving from hunting activities were supplemented by substantial losses from predators that weakened the stabilizing effect of a high birthrate in maintaining the species population.

Certain other implications of the calculation procedure mentioned above should also be noted.

The rate of decline in numbers of hunted animals was initially slow and gradually increased. Towards the end herds that were still relatively large became extinct rather quickly. Beyond this, long before the full extinction of a particular animal took place, while its numbers were still declining, prolonged periods of total absence of such animals arose in particular regions as a result of natural fluctuations in their numbers and the migration of individual herds.

Thus a threat arose with increasing frequency to the growing population of Europe during the Paleolithic that was primarily due to mass hunting activities. This explains the intense interest of men from the Upper Paleolithic in making drawings of animals that were attributed magic properties and were intended to help restore the vanishing herds. Yet continued intensive hunting activities led to the destruction of large herbivores at an increasingly rapid rate, and since its last stage was particularly rapid, this placed hunting tribes in a difficult situation in which they were not given enough time for a gradual transition to other forms of acquiring food. It may therefore be thought that the transition to the Mesolithic was difficult for primitive societies, and was accompanied by temporary reductions in their population.

It is known that the end of the Paleolithic at middle latitudes of North America was also associated with the extinction of large animals, and this confirms the nature of the corresponding causal relations. At the same time it is quite probable that perceptible changes in natural conditions towards the end of the Paleolithic produced additional difficulties for the continued existence of these animals and served to intensify a process which was already taking place.

Thus the termination of the Paleolithic culture in Europe may have partly been the outcome of an irreconcilable contradiction between the mass hunting technology that was created by man in the Upper Paleolithic which produced a temporary abundance of food and made possible the increase in population, on the one hand, and the limited nature of natural resources for those hunting activities which became depleted after a certain period of time on the other.

But the following question then arises: What brought about a disruption of the balanced ecology of the pre-Holocene biosphere when the species of animals pursued by predators could exist over hundreds of thousands and even millions of years?

It is possible that the major reason for such a disruption was an extremely high rate of evolution in man by comparison with the rate of evolution of the animals that he was pursuing.

As has already been noted, large animals possessing a low birthrate are less able to maintain a stable population when hunted by predators than are small animals possessing a higher birthrate. Accordingly, the large size of their bodies gives such animals advantages in struggling for existence only in those cases in which they are relatively secure from attacks by predators. If, on the other hand, their size does not protect them from attacks (as was the case even with mammoths in the Upper Paleolithic) the only way in which populations could be maintained is an increase in the birthrate which in the case of mammals is usually achieved through a reduction in the size of their bodies.

Such a change on the part of the animals being hunted, given the rate of their evolution in the Pleistocene, required a much larger length of time than the period during which their extermination occurred.

A similar approach was recently applied by P. Martin (1973) in clarifying the reasons for the vanishing of many types of large animals in North and South America towards the end of the Pleistocene. It is known that man appeared on these continents relatively late, namely, towards the end of the Würmian glaciation, and that the regularities governing the process of his settlement in America are still poorly understood. Martin has assumed that the first small group of persons crossed the ice-bound Behring Strait and entered the north-western regions of North America where they found very numerous herds of large animals that had never encountered man and did not possess instincts favouring their protection when attacked by man. This produced more favourable conditions for hunting in America than in the countries of Europe and Asia, where wild animals were in contact with man at the various stages of his evolution over periods of hundreds of thousands of years and were best able to protect themselves.

In such conditions a small group of first inhabitants of America could supply itself with an unlimited quantity of food products, and, having rapidly increased in numbers, settle all of America, destroying on its way most large animals it encountered. Martin's calculations show that over a period of a thousand years the initial wave of settlers could reach Tierra del Fuego—the southern most tip of South America, having destroyed on its way approximately thirty genera of large animals that existed in America towards the end of the Pleistocene (the mammoth, the mastodon, the megathere, horse, camel, and many others). Following their extinction hunting tribes were forced to shift to other sources of food.

The conclusion that the first culture of modern man—the Upper Paleolithic—led to the emergence of an ecological crisis on large territories that destroyed that culture may be related to the well-known statement by Marx: "Cultivation when it progresses spontaneously and is not consciously controlled leaves deserts behind it." *

Man's subsequent history has produced much data that confirms the difficulties of maintaining an ecologically balanced biosphere under conditions of spontaneously developing economic activities.

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* K. Marx and F. Engels, *Selected Correspondence*, Moscow, 1965, p. 202.

IX

Man and the Biosphere

1. The Current Ecological Situation

Anthropogenic Ecological Crises of the Past. Beginning with the Upper Paleolithic man's economic activities have often led to a worsening of natural conditions and this has created varying difficulties for the further existence of human society. Many ecological crises developed in this way that spread to territories varying in size.

From the very earliest times an important factor relating to man's influence on the environment was fire, whose application made it possible to destroy vegetation over large areas. Forest fires and steppe fires have long been used in hunting large animals. Such a method was still used in recent times by Australian aborigines, who destroyed vegetation on territories covering tens of square kilometers (J. Dorst, 1965). Similar hunting methods were probably used by hunters of the Upper Paleolithic.

It is clear that such fires over large territories led to the destruction of wild animals and of natural ecological systems. At the same time the destruction of forests facilitated further hunting activities in relation to large animals, and this perhaps explains the rapid destruction of forest vegetation in many regions following the appearance of modern man.

During the Neolithic, when the principal economic activity was the raising of herds and agriculture, the burning of plant covers assumed very large proportions. It was applied in order to expand pastures by destroying forests and especially in the slash-and-burn agricultural practices. The resulting fertilisation of the soil by ashes produces abundant harvests, even when the plowing is not very deep.

Under such a system of agriculture the fertility of soils declines rapidly, new forest sectors must be cut after a few years (sometimes after one-two years) and agricultural activities must then be moved to these new sectors. Such a method may be applied when there exist large sparsely-settled forest territories. In the past it was widely used in many countries at middle latitudes. and today is employed in a number of developing countries in tropical areas.

The wide use of such methods of burning vegetation over large territories has led to sharp changes in natural conditions, including the flora, fauna, soil, and possibly the climate and hydrological regime as well. Since the systematic burning of vegetation, both at middle latitudes and in tropical regions was initiated a very long time ago, it is difficult to estimate the full scope of the environmental changes that this has produced. Observations show, moreover, that in many cases the plant cover destroyed by man does not restore itself even after the cessation of its systematic burning.

Aside from the impact of this type of agriculture in a number of regions forests were also destroyed to produce timber. Beyond this the grazing of agricultural animals has produced a major impact on the natural plant cover of many regions. Such grazing activities were frequently carried out without considering the need to restore the plant cover. In forest regions possessing a dry climate the feeding of goats and other animals on the leaves of young trees eventually led to the destruction of forests. Excessive grazing of cattle destroyed the vegetation of dry steppes and savannas that subsequently frequently turned to semi-deserts and desert areas.

The influence of economic activities on plant covers appears to have frequently produced severe losses to human society.

One of the first civilizations in human history appeared in northwest India (the third-second millennium B.C.). The centers of that civilization (Harappa, Mohenjo-Daro, and others) are found in regions that have since become deserts. Some authors believe that in the past these regions were dry steppes possessing favourable conditions for animal husbandry and certain types of agriculture.

An excessive grazing of cattle could have resulted in the destruction of plant cover, producing an increased temperature and a decline in the relative humidity of the lower air layer. As a result the amount of precipitation decreased, and this made impossible the renewal of the plant cover. It is thus possible that anthropogenic changes in climate were one of the causes that led to the vanishing of India's ancient civilization.

Another example relates to changes in natural conditions in the basin of the Mediterranean Sea in ancient times. Originally there existed large forests on the territory of Greece and of a number of other Mediterranean countries that were subsequently cut and partly destroyed as a result of excessive grazing of cattle. This served to intensify erosion and led to the complete destruction of the soil cover on many mountain slopes. This, in turn, intensified the climate's aridity and produced a major deterioration of conditions of agricultural production. In this particular case changes in natural conditions did not cause substantial losses to ancient civilizations, but they did influence many aspects of human life in ancient times.

The view has been expressed that the depletion of soils on territories occupied by the Maya State in Central America through the burning of forests for agriculture was one of the reasons for the destruction of that highly-developed civilization. Europeans who settled in Central America found numerous cities in the tropical jungles that had long been abandoned by their inhabitants.

There are many other such examples relating to the creation of adverse environmental conditions by man, that were irreversible in the light of contemporary technical possibilities. Such cases may be described as anthropogenic ecological crises.

In studying anthropogenic ecological crises of the past one may advance the hypothesis that adverse conditions were particularly serious in the case of crises that developed at the early stages of human society. This is particularly true of the crisis during the Upper Paleolithic that we have described. Later ecological crises did not extend to such large areas and appeared to have produced consequences that were less lasting.

It should be noted that the influence of ecological crises on the development of human society was ambivalent. Together with the damage that they caused to man's traditional economic activities the crises stimulated development of new ways for meeting the economic requirements of society. In particular, this refers to the crisis of the Upper Paleolithic, namely disruption of hunting large animals. Eventually this served to intensify the emergence of new and far more effective types of economic activities, namely the raising of cattle and agriculture.

The substantial deterioration of conditions of agriculture in a number of ancient Mediterranean countries that has been mentioned exerted a definite influence on the development of seafaring, crafts, and other types of non-agricultural production. It must be supposed, however, that a stimulating influence of ecological crises on economic progress was not always present, and that in any event it played a smaller role than did the social development of society.

The history of man's influence on the biosphere shows that technical progress constantly increased his possibilities for influencing the environment and created conditions for the emergence of major ecological crises. On the other hand, this progress extended the possibilities for overcoming deteriorations in the natural environment that were produced by man. These two opposing tendencies have been particularly pronounced in the second half of the twentieth century.

Man's Influence on the Biosphere. In our present age human activities exert an enormous influence on the natural conditions of the entire planet.

Changes in the flora and fauna of land areas are particularly pronounced. Many types of animals have been completely destroyed by man and a still greater number of species are confronted with extinction. It is believed that in recent times more than 120 species and subspecies of mammals have vanished as well as 150 species of birds (J. Dorst, 1965).

Plant covers have experienced enormous changes over a large part of the surface of continents. Wild plants have been destroyed over large areas and have been replaced by agricultural fields, while those forests that have been preserved are largely secondary, in the sense that, by comparison with natural plant covers, they have been greatly altered as a result of human activities. Similarly, major changes have occurred in the plant cover of many steppe and savanna regions as a result of the intensive grazing of domesticated animals.

Man's influence on natural plant covers has also influenced the process of soil formation in the corresponding regions, and has led to changes in the physical and chemical properties of soils. Even greater changes have been experienced by soils in agricultural areas as a result of their systematic utilization, the application of fertilizers, and the removal of a substantial part of the biomass of corresponding plants. In many regions the cultivation of soil has led to an increased rate of erosion, and as a result soil covers have become destroyed over large areas.

The influence of human activities on the hydrological regime of land areas has been increasing rapidly. The run-off of many large rivers as well as of small ones has been substantially altered as a result of the creation of hydroengineering facilities. A substantial part of the waters of river run-off is diverted in order to meet the needs of industry and of urban populations and for irrigating agricultural lands. The creation of large reservoirs whose area is often comparable to that of large natural lakes greatly alters the evaporation regime and the run-off regime on large territories.

The pollution by man of the atmosphere and of the waters of continents and of oceans is proceeding on an increasing scale.

It has already been noted that, even though the activities of modern man have greatly altered the natural environment of our planet, these changes usually represent a sum of local influences on environmental processes. It is not as a result of changes by man of global natural processes that they acquire a global character but rather because local (or else regional) influences are disseminated over large areas.

It is only the type of human influence on climate that is examined in the next section of this chapter that holds a special position in this respect.

Although many of the changes in natural conditions that have been mentioned exert an unfavourable influence on human life and activities, the question whether such changes may be viewed as manifestations of an ecological crisis is far from being a simple one.

Earlier we have defined an ecological crisis for the age of human society as an irreversible deterioration in man's environment, that is, as a change in that environment that cannot be overcome by man in the course of a reasonably short period. The examples of ecological crises of the past that were mentioned earlier are of such a type.

The question of the extent to which current unfavourable changes in natural conditions are irreversible calls for close attention. Leaving aside for the moment anthropogenic changes in climate let us note that the experience of the past few decades shows that in principle it is possible by relying on modern achievements in science and technology to overcome nearly all the recent changes in the natural environment that are unfavourable to man.

The restoration of forests and of other types of natural plant cover is widely carried out in many countries. Sometimes that problem is met with the help of simple means, as, for example, through limitations on cattle grazing. In other cases more expensive measures are needed that are nevertheless generally accessible, even for countries whose economic development is relatively low.

Much experience has been accumulated in preserving and restoring soil covers and in limiting the destructive effects of erosion. The protection of air and of water bodies from pollution and the purification of polluted water bodies often require large expenditures. But available data show that it is possible to restore even very large water bodies, such as, for example, some of the Great Lakes in the United States, through expenditures that represent a relatively modest share of the country's state budget.

Significant results have been achieved in a number of countries in preserving animal species threatened with extinction and in increasing the population of endangered species. Such animals include sea otters, fur-seals, bisons, beavers, sables and many others.

At the same time measures intended to preserve wild animals do not always produce positive results. This deserves close attention, since of all types of harm caused by man to environment the destruction of the genetic fund of currently existing organisms is the only irreparable one.

It has already been noted that man has already destroyed hundreds of animal species and subspecies. This may appear to be not very important, since the number of lost species represents a negligible part of all existing species, that were not significant, moreover, from the economic point of view.

But such considerations are completely misleading. The significance of a particular animal species for mankind is not limited to current economic estimates, and the loss of any part of the current genetic fund may deprive man in the future of important means for regulating ecological systems and of possibilities for a comprehensive utilization of living nature for various practical objectives.

It should not be concluded that all forms of local changes in the environment produced by man (with the exception of the destruction of certain animal species) are irreparable. The possibility of overcoming nearly completely the harm that has been done to nature on our planet makes it possible to hold that there does not exist a global ecological crisis at the present time.

But there are many local ecological problems which can be settled with the aid of modern scientific and technical achievements, but the probability of their settlement in the near future is not always manifest. Thus, for example, it is often impossible for capitalist countries that do not possess the needed systems of centralized planning to find any means required for restoring favourable natural conditions. This leads to an intensification of local processes of environmental disruption and transforms them into elements of regional ecological crises.

Thus at the present time the problem of ecological crises is primarily a social problem that is associated with the character of social development in individual countries.

Forecasts of a Global Ecological Crisis. Quite aside from any particular evaluation of the current ecological situation the view has been often repeated in recent years that a global ecological crisis may develop in the relatively near future as a result of which a general economic crisis will take place and the world's population will decline sharply. The most detailed analysis of such a point of view has been presented in a study by J. Forrester (1971), and in the collective monograph *Limits to Growth* (D. Meadows et al., 1972).

As did the studies concerned with the economy of the Upper Paleolithic presented in the preceding chapter (Budyko, 1967), Forrester and the authors of *Limits to Growth* solved equations determing increments in population in the context of other relations characterizing the depletion of resources. As in the case of calculations pertaining to the Upper Paleolithic the analysis carried out by these scientists in relation to the present age has led to the conclusion that a further growth in population will produce a crisis that will lead to a sharp reduction in population. The estimated time for the emergence of such a crisis is quite short—approximately one hundred years.

It should be noted that the numerical models employed in these studies are more complex than the relatively simple model that was used in our own study. The systems of equations that were employed can only be solved with the help of computers.

Since in the model relating to the Upper Paleolithic it was possible to consider food resources alone (mineral resources did not influence population at that time, while pollution did not yet exist) the calculations relating to that model could be carried analytically. In models relating to the current age, the influence of a number of factors on changes in the population was taken into account, including the volume of food, mineral resources, and environmental pollution.

Let us consider more closely the results of the forecast of changes in economic conditions presented in *Limits to Growth* that were obtained by applying Forrester's model.

The corresponding calculations were carried out for the period from 1900 to 2100. They expressed the dependence of birthrates and deathrates on a variety of factors reflected in empirical data. In cases in which such data were insufficient (for example, in estimating the influence of environmental pollution on the deathrate) the authors of the study relied on hypothetically formulated relationships.

The results of the calculation show the industrial production and the production of foodstuffs per capita increase until the beginning of the twenty-first century and after that both these magnitudes decline rapidly to values that are far below their levels at the beginning of the twentieth century.

Reserves of mineral resources at the beginning of the twenty-first century decline sharply and by the middle of the century are practically depleted. Environmental pollution increases rapidly during the first half of the twenty-first century but then declines again by the middle of the century as a result of declining industrial production.

Population increases to the middle of the twenty-first century and then begins to decline and towards the end of that century is reduced by nearly one-half. In this case the principal cause for the crisis is the depletion of mineral resources. This produces a sharp increase in mortality because of a reduction in the output of food products and a worsening of medical services to the population.

The results of this calculation are supplemented in *Limits to Growth* with a number of other calculations relying on assumptions. In most cases, however, these assumptions do not alter the major conclusion in the initial study. For example, the assumption that known reserves of mineral resources will be doubled produces a still sharper decline in the population by the middle of the twenty-first century, since this results in a catastrophic increase in the level of environmental pollution.

The assumption that effective control is introduced with regard to environmental pollution and that reserves of mineral resources are doubled does not prevent a decline in population attributable to shortages of food products.

Additional calculations show that in order to prevent a reduction in the population and maintain a high level of industrial production a combination of measures is required that includes a cessation in population growth, a sharp reduction in the use of mineral resources, a cessation of growth in industrial production and a shift of efforts to overcoming the effects of environmental pollution as well as to increasing the output of food products.

It also follows from these calculations that in order to stabilize economic conditions in the twenty-first century such measures must be introduced immediately, and that their introduction towards the end of the twentieth century will be too late and will not forestall a crisis. The authors observe that the probability that such measures will in fact be introduced in time is not very great, and expect that an ecological crisis will be followed by "social disintegration" that is, by a disintegration of the existing organization of society.

A critical evaluation of Forrester's studies and of those of the authors of *Limits to Growth* is given by Ye. K. Fyodorov (1972), who lists a number of fundamental shortcomings.

These include a failure to take into consideration those fundamental changes that are introduced by technological progress into a society's economic development and also the manner in which man's needs are met.

At the same time Fyodorov notes that the conclusions of Forrester and of the authors of *Limits to Growth* concerning the forthcoming crisis in society's interaction with nature reflect correctly the principles that govern the spontaneous development of bourgeois societies. The basic socio-economic characteristics of the capitalist system hinder a rational utilization of natural resources, inhibit the solution of the problem of supplying food, and make difficult a reliance on planned measures in struggling against environmental pollution. In such conditions there emerges the possibility of a crisis in man's interactions with nature, and many traits of that crisis may already be clearly seen in the present age. Fyodorov also notes that it is possible to overcome the obstacles that inhibit man's further technological and economic development only through the institutions of a socialist planned economy.

The conclusion that long-term economic planning is needed is contained in *Limits to Growth*, but the set of planning measures that is proposed in order to forestall a crisis are clearly of a utopian character.

One would expect that even within the framework of the limitations that attach to the numerical model employed in that study it is possible to find more realistic ways for overcoming the difficulties that the authors note. For example, simple calculations show that comprehensive measures directed against environmental pollution, together with the mass use of synthetic materials and even modest further achievements in the chemical synthesis of food products would largely remove anticipated obstacles to technological and economic development.

In noting the substantial limitations that attach to currently existing methods of modeling economic development, the possibility of improving these methods should be studied so that they may be applied to specific problems that arise in the long-term planning of technological progress.

The use of numerical models that permit calculating the probable changes in population as related to basic economic indicators and to the nature of changes in environmental conditions makes it possible to formulate well ahead of time the technological problems that must be solved in order to overcome the various obstacles that confront economic development.

One such major problem that is not examined in the studies of the American authors concerns the possibility of changes in our planet's climate as a result of economic activities. The exceptional importance of that problem stems from the fact that in the light of available data such a change may exert a major influence on economic activities before the other global ecological disruptions that are examined in Forrester's study and by the authors of *Limits to Growth*.

2. *Man's Influence on Global Processes*

The Climate of Cities. Among the various components of the biosphere the atmosphere is best able to carry the disturbances that arise within it over long distances. Atmospheric processes represent the major mechanism that may transform local influences of man on his environment into global changes in natural conditions.

Man's influence on the atmosphere began to manifest itself thousands of years ago. The destruction of forests in order to till the land, the creation of a system of artificial irrigation facilities and the draining of swamps have all produced certain changes in local climatic conditions. To some extent global climate may have been influenced by the burning of vegetation over large areas, which increased the concentration of carbon dioxide and the amount of aerosol particles in the atmosphere. But that influence was relatively negligible. Man's influence on climate has increased in the age of industrial development, and especially over the past few decades.

Some notion concerning the nature of the influences of human activities on climate may be derived from meteorological observations in cities. These observations show that climatic conditions in cities usually change perceptibly by comparison with outlying regions, and that, other conditions being equal, these changes are greater as the size of city territory increases.

There are data indicating that changes in the climate of large cities appeared centuries ago. In particular, H. Landsberg (1956) cites the observations of a resident of London in the seventeenth century concerning the pollution of that city's air. This reduced solar radiation in the city substantially by comparison with rural localities.

The major factors influencing the meteorological regimes of cities include:

1) Changes in the albedo of the Earth's surface, which is less for regions in which construction has taken place than the albedo of non-urban regions;

2) Changes in the average evaporation from the Earth's surface, which is much lower within city boundaries (even though immediately after precipitation evaporation from roofs and streets may be greater than evaporation in non-urban localities);

3) The release of heat produced by various types of economic activities in amounts that are comparable to the quantity of solar energy reaching the city's territory;

4) An increase in the roughness of the Earth's surface by comparison with non-urban localities;

5) A pollution of the atmosphere by a variety of solid and gaseous mixtures formed as a by-product of economic activities.

One of the major features of urban climates concerns the formation of "islands of heat" within cities that are characterized by higher temperatures than in localities outside the city. That effect has been examined in many experimental studies, and a number of numerical models have been proposed.

The findings of such studies (*Inadvertent Climate Modification*, 1971; H. Landsberg, 1970, 1972; M. Berlyand and K. Kondratyev, 1972) show that such "islands of heat" usually possess a complex structure, and that each urban block operates as a source of heat for those areas around it that are free from buildings. The average air temperature in a large city often exceeds that of surrounding districts by 1-2°C, and at night, in the presence of a moderate wind, the difference in temperature may attain 6-8°C. This is reduced as winds become stronger.

Interesting observations describing the formation of "islands of heat" as urban construction proceeds have been presented by H. Landsberg and T. Maisel (1972). In the United

States construction activities intended to develop a new city called Columbia were initiated near Washington in 1967. In 1968, when only a small part of that city had been constructed, the air temperatures at its center in the evening was approximately 0.5°C higher than in outlying localities. In 1970, following a substantial increase in construction, temperature increases in the city's center attained 4.5°C, and was 2°C over a large part of its territory. In the daytime, whenever the intensity of air movements increased, increases in temperature were perceptibly lower.

At the same time as increases in temperature occurred on the territory of Columbia the relative air humidity declined by several per cent. This appears to have been influenced by both the increasing temperature and a reduced rate of evaporation.

It is evident that "islands of heat" appear largely as a result of the influence of the first three of the factors that have been listed that determine urban climate. The relative importance of each of these factors in different cities and in different seasons may vary considerably.

It should be noted that not only the relative but also the absolute air humidity is usually reduced within "islands of heat", because of the reduced evaporation in areas on which buildings have been erected. A further heating of the air over cities produces local circulation systems similar to breezes and also intensifies upward convective motions over cities. At the same time, because of the influence of buildings, wind velocity is perceptibly lower in cities.

Various types of air pollution are relatively high in many cities, and it is that feature of their climate which is of greatest practical concern. Air pollution originates in by-products of enterprises, heating systems and transportation systems. A substantial amount of anthropogenic aerosol in cities is produced from discharges of solid or liquid particles, or else originates by way of transformation of gases emitted into the atmosphere.

Increases in the concentration of aerosols over cities sharply reduce the amount of solar radiation reaching the Earth's surface. According to data prepared by Landsberg (1970) direct solar radiation in large cities is frequently reduced by approximately 15 per cent, ultraviolet radiation by an average of 30 per cent (in winter months it may vanish altogether), and the duration of sunshine is reduced by 5-15 per cent. A major role in reducing solar radiation in cities is played by a thin lower air layer that contains most of the aerosol particles. Within that layer horizontal visibility is usually sharply reduced and is often 10-20 per cent of its value outside city boundaries.

A high concentration of aerosol particles in the air over cities contributes to frequent fogs, including especially stable fogs of the smog type whose small drops contain a significant volume of pollutants. In some cities, such as Los Angeles, where local conditions contribute to a weakening of atmospheric circulation, smog may last many days and is a source of considerable harm to the population's health.

An increase in the number of condensation nuclei in the air over cities and an intensification of updrafts of air leads to an increase in cloudiness and precipitation. There are data that indicate that in certain industrial centers there exists a weekly cycle in the quantity of precipitation that is explained by a reduction in precipitation on weekends, when industrial enterprises are closed. A certain increase in precipitation and a significant decline in evaporation produce water run-offs and in many cities this is assisted by special systems for removing waters originating in rainfall. In cities with heavy snowfalls snow is often removed beyond city boundaries. In the spring this produces higher temperatures within the city by comparison with outlying localities.

The climate of cities may be greatly improved through rational location patterns for residential housing and productive facilities, and through the planting of greenery and measures for reducing air pollution.

There are many examples illustrating how changes in heating systems (shifts from solid fuels to natural gas or else electrical energy) have sharply reduced air pollution in cities and, hence, the related changes in various elements of climate. The location of industrial enterprises that pollute the atmosphere outside city boundaries and the application of effective means for cleaning the stack discharges and other pollutants are equally important, as is the creation of large parks, and the planting of greenery along streets and in various other areas.

One should note that as urbanization proceeds in very densely populated regions certain traits of urban climates are spread to territories that may stretch over hundreds of kilometers. In such cases there occurs an aggregation of the influences on climate of numerous areas of settlement that are located close to each other, each of which operates as a source of heat and air pollution. Such situations exist in a number of regions in the USA, Japan, and Western Europe, where local changes in climate encompass increasingly wider areas.

Changes in Global Climate. In recent years data have been produced showing that man's current economic activities influence not only the local climatic conditions for individual regions but also the climate of our planet as a whole. One of the factors in such an influence concerns changes in the volume of carbon dioxide in the atmosphere.

The hypothesis that as a result of burning large quantities of coal, petroleum and other types of fuel the quantity of carbon dioxide in the atmosphere has begun to increase was already formulated in the first half of our century (G. Callender, 1938). It was only in the 1950s, however, in connection with the activities of the International Geophysical Year, that systematic observations of atmospheric carbon dioxide were carried out at a number of stations and this made possible a quantitative estimate of the growth in the concentration of carbon dioxide.

These observations have shown that together with a perceptible yearly cycle of concentration of carbon dioxide at the Earth's surface (a reduction in concentration during the summer in connection with increased photosynthesis) there exists a clear trend towards a growing concentration from year to year. Observations carried out at Mauna Loa (Hawaiian Islands) for the years 1958 to 1968 have shown that the yearly increase in concentration of carbon dioxide was $0.64 \cdot 10^{-4}$ per cent, which represents a substantial part of the volume of carbon dioxide that enters the atmosphere each year as a result of the burning of various types of fuel (*Inadvertent Climate Modification*, 1971). The findings produced by the observations at Mauna Loa accord well with data from similar observations in Alaska, Sweden and in Antarctica (B. Bolin and W. Bischof, 1970, etc.). Since these observations were carried out at points on the Earth's surface that are very distant from each other, there are no grounds for doubt that they correctly reflect an existing tendency towards a change in the concentration of atmospheric carbon dioxide.

It follows from the observations that approximately one-half of the carbon dioxide that results from human activities is retained in the atmosphere. The second half appears to be absorbed by the ocean and to a lesser extent by living organisms.

The problem concerning the mechanism through which additional carbon dioxide is absorbed by water bodies and living organisms is poorly understood from the quantitative point of view. While oceans possess a large potential capacity in this regard and may absorb an enormous quantity of carbon dioxide, that rate of absorption is greatly reduced because of a slow exchange between surface layers and deeper layers of ocean waters.

As the concentration of carbon dioxide increases so does the rate of photosynthesis. Yet the additional quantity of organic substance created in this way is mineralized after a certain time and releases the carbon dioxide that was expended on its creation.

The development of a comprehensive quantitative theory making possible to consider the influence of buffer processes in the ocean and the biosphere on changes in the concentration of carbon dioxide in the atmosphere is a task for the future. At the present time simplified procedures may be employed towards this end such as the simple numeric model proposed by L. Machta (*Inadvertent Climate Modification*, 1971). It follows from calculations based on that model that during the past one hundred years the quantity of carbon dioxide in the atmosphere has increased by 0.003 per cent, i.e., approximately by 10 per cent of its current concentration and that the major part of this increase in content of carbon dioxide has occurred in recent decades.

In view of the findings of calculations relating to the influence of the concentration of carbon dioxide on air temperature at the Earth's surface it is possible to establish that the increase in the concentration of carbon dioxide attributable to industrial activities has raised the average global temperature at the Earth's surface by 0.3-0.4°C. This magnitude is comparable to the temperature fluctuations that have occurred during the past century. Thus it is probable that the burning of various types of fuel produces a definite influence on current climatic conditions.

The authors of many studies have come to the conclusion that as a result of economic activities a large quantity of particles enters the atmosphere which increases appreciably the concentration of atmospheric aerosols.

Observations of changes in the dust content of the atmosphere have been carried out by F. F. Davitaya (1965), who relied on data concerning the vertical distribution of dust concentration in the snow cover of glaciers in the Caucasus. These data have shown that the quantity of dust per unit volume in the upper layer of the snow has increased substantially by comparison with the deeper layer that was formed earlier. In the opinion of Davitaya this difference corresponds to a sharp increase in the concentration of dust in the atmosphere that has occurred in recent decades.

The studies of R. McCormic and J. Ludwig (1967) and of other authors cite data showing that there has occurred a reduction in the transparency of the atmosphere under clear sky conditions during the past several decades that appears to be attributable to a growing concentration of atmospheric aerosols.

Available estimates indicate (*Inadvertent Climate Modification*, 1971) that the volume of anthropogenic aerosols that enter the atmosphere each year at the present time is approximately 200-400 million tonnes, which is 10-20 per cent of the overall volume of aerosol particles entering the atmosphere. Only a minor part of the total volume of anthropogenic aerosols enters the atmosphere in the form of solid or liquid particles. Their main source lies in gaseous admixtures emitted by industrial plants such as sulphur dioxide and nitrogen oxides that produce aerosol particles through various chemical reactions.

The data derived from observations which are discussed below show that anthropogenic aerosols increase substantially the overall concentration of aerosol particles not only in cities and individual industrial regions but over large areas as well, and that this lends a global character to the current process of air pollution.

In Chapter III the influence of stratospheric aerosols on the atmospheric thermal regime at the Earth's surface was noted. It was already shown by W. Humphreys (1929) that

that influence was largely produced by a reduction in the flux of short-wave radiation entering the troposphere. Aerosol particles in the stratosphere played the role of screen, which in some measure alter the meteorological solar constant, i.e., the amount of solar energy reaching the upper boundary of the troposphere. The absorption of radiation by these particles may cause local heatings of the stratosphere; such heatings, however, exert a little influence on the thermal regime at the Earth's surface because of the insignificant air density in the stratosphere. Thus, as a series of studies show, increases in the concentration of aerosols in the stratosphere inevitably bring about a reduction in temperature at the Earth's surface.

A more complex influence on the thermal regime is produced by aerosol particles in the troposphere. These particles attenuate the flux of short-wave radiation reaching the Earth's surface as a result of a back scattering and an absorption of radiation by aerosol particles. While the first of these processes increases the albedo of the Earth-atmosphere system, the second one reduces it. The influence of tropospheric aerosols on the thermal regime is largely determined by the relation between the albedo of the Earth-atmosphere system and aerosol concentration. Increases in albedo produce a declining temperature at the Earth's surface, while declines in its magnitude lead to increased temperatures.

It is clear that changes in the reflectivity of the Earth-atmosphere system under the influence of aerosols will depend on the albedo of the Earth's surface. The lower that albedo, the more probable it is that atmospheric aerosols will increase the reflectivity of the system. For large values of the albedo at the Earth's surface (snow, ice) the probability of a reduction in the albedo of the Earth-atmosphere system under the influence of aerosols increases.

Empirical data show that in the presence of cloudiness the influence of tropospheric aerosols on the thermal regime of the atmosphere is apparently not very great.

One may arrive at a similar conclusion by considering that the major volume of tropospheric aerosols is found below the upper boundary of clouds as a result of which the influence of aerosols on the radiation regime and the thermal regime in the presence of cloudiness is substantially weakened. If one assumes that on the average approximately one-half of the Earth's surface is covered with clouds, it may be expected that the overall influence of tropospheric aerosols on average air temperature is equivalent to a reduction in that temperature by approximately one-half of the reduction that would exist in the absence of clouds in the atmosphere.

In estimating the influence of anthropogenic aerosols on climate, use was made of data from observations of groups of actinometric stations located in Europe, Asia and America. This made it possible to construct a graph of secular variations of direct solar radiation under clear sky conditions over the past few decades (Budyko, Vinnikov, 1973).

Figure 34 represents the magnitudes of anomalies in direct radiation expressed as percentage deviation from the normal over a period ranging from 1885 to 1970. Curve *A* in that Figure represents values that have been averaged over running five-year periods.

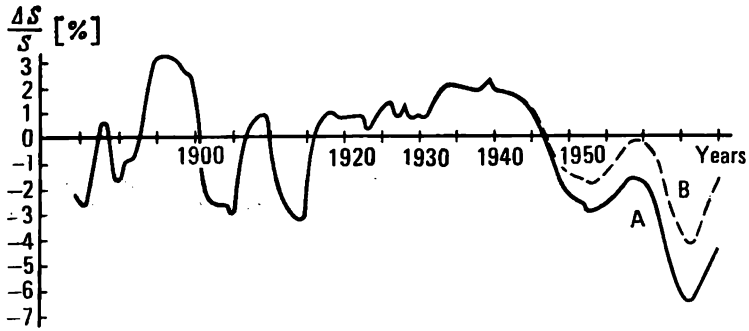

Fig. 34. Secular Variations of Direct Radiation

It may be seen from that Figure that from the end of the nineteenth century until the 1920s several sharp declines had occurred in the values of direct radiation that are explained by the influence of large volcanic eruptions—in Krakatau, Mont-Pelée, Katmai, and others after which the concentration of aerosols in the lower layers of the atmosphere increased substantially. Subsequently until the 1940s direct radiation attained maximal values and changed relatively little before it began to gradually decline.

A number of authors have expressed the view that the gradual decline in direct radiation since the 1940s is explained by increases in the volume of atmospheric aerosols produced as a result of human activities.

It should be noted that anthropogenic aerosols are largely concentrated within the lower troposphere. Accordingly, their influence on the radiation reaching the Earth's surface is substantially less in the presence of clouds than under clear sky conditions. It has already been mentioned that there exists a definite relationship between changes in direct radiation resulting from its scattering on aerosol particles and corresponding changes in total radiation, and that data concerning such changes make it possible to calculate changes in total radiation from fluctuations in direct radiation.

If these fluctuations occur as a result of the scattering of radiation on aerosol particles in the stratosphere the corresponding calculations may be based on data concerning changes in direct radiation shown by curve *A* in Figure 34. If, on the other hand, a part of these fluctuations is produced by a reduction of radiation in the lower troposphere, calculations of average global changes in total radiation should take into consideration the influence of clouds on the atmospheric radiation regime.

Let us assume that at the beginning of the 1940s changes in direct radiation were attributable to two processes, namely an increase in the volume of anthropogenic aerosols which gradually reduced the quantity of radiation, and fluctuations in the transparency of the stratosphere, which could reduce direct radiation when the concentration of aerosols in the stratosphere became greater or else increase it to a maximum value corresponding to full or nearly full absence of aerosols in the stratosphere.

Suppose also that changes in direct radiation attributable to increases in anthropogenic aerosols are described by a straight line passing through the points on the curve that correspond to a maximum transparency of the atmosphere in individual years. It may be presumed that for increased values of transparency in the atmosphere the stratosphere is either completely or nearly completely freed from aerosol particles. This is confirmed by data from direct observations of stratospheric aerosols. The resulting estimate of reduction in the radiation through anthropogenic aerosols will correspond to conditions of cloudless skies. Let us assume that the average cloudiness is 0.5 for the Earth as a whole, and let us also assume that in the presence of clouds the influence of anthropogenic aerosols on total radiation is small by comparison with their influence in cloudless skies. We will then find that in calculating the influence of changes in radiation on air temperature one should make use of data concerning values of direct radiation that are increased by one-half of the changes in direct radiation under cloudless skies attributable to the influence of anthropogenic aerosols. The curve representing direct radiation embodying such an adjustment is represented in Figure 34 through the broken line *B*.

Let us note that data relating to changes in direct radiation attributable to increases in the mass of anthropogenic aerosols make it possible to arrive at the approximate estimate of that mass. It may be seen from that Figure that in recent years anthropogenic aerosols have produced a decline in the direct radiation under cloudless skies by approximately 6 per cent. For average conditions of cloudless skies this corresponds to a reduction in total radiation by approximately 1 per cent. The mass of aerosol that changes (as a result of diffusion by its particles) the total radiation by 1 per cent will be, according to calculations based on formulae of atmospheric optics, approximately 10^{-6} g/cm^2. Thus the quantity of anthropogenic aerosol attains approximately 5 million tonnes for the entire globe.

That figure corresponds to the volume of optically active aerosols for which the particle sizes are equal to tenths of a micron. It is usually considered that approximately one-half of the volume of aerosols in the troposphere consists of the so-called "gigantic" particles, whose number is negligible by comparison with particles possessing the dimensions noted above, and which therefore do not exert a substantial influence in attenuating radiation. Taking this into account we will find that the overall volume of anthropogenic aerosols may be approximately 10 million tonnes. Since in the Southern Hemisphere the volume of anthropogenic aerosols is substantially less than in the Northern Hemisphere (the data mentioned above concerning the reduction in direct radiation refers to the Northern Hemisphere) it may be presumed that the global volume of anthropogenic aerosols is equal to 5-10 million tonnes.

Let us compare that figure with the estimate of the volume of anthropogenic aerosols obtained through direct methods. According to current data (SCEP, 1970), a volume ranging between 1,000 and 2,600 million tonnes of matter enters the troposphere each year and forms aerosol particles, and 10 to 20 per cent of that volume is attributable to human activities. Assuming that the average length of life of particles in the troposphere is approximately 10 days we will find that the total volume of aerosols in the troposphere is 30-70 million tonnes, while the volume of anthropogenic aerosols is 3-14 million tonnes. The earlier estimate of the mass of anthropogenic aerosols accords well with these figures, and this confirms that the corresponding calculations of the influence of human activity on the secular variations of direct radiation are sufficiently accurate.

Chapter III contains calculations relating to changes at middle latitudes in air temperature for the period of 1910 to 1950 based on fluctuations in solar radiation. In establishing the causes of changes in air temperature in later years it is necessary to take into consideration the influence of human activities on climate. It has been noted earlier that over the last few decades changes in direct radiation under cloudless skies have depended not only on fluctuations in the concentration of aerosols in the stratosphere but also on changes in the quantity of aerosols in the troposphere resulting from human activities. The data presented in Figure 34 show that the influence of changes in the quantity of tropospheric aerosols on radiation has gradually increased and at the present time exceeds the influence of changes in the volume of aerosols in the stratosphere.

In recent decades air temperature at the Earth's surface has also been influenced by increases in the concentration of carbon dioxide in the atmosphere, which may raise the average temperature of the Earth's surface by approximately 0.3-0.4°C.

Since according to data based on observations the average air temperature in the 1940s-1960s in the Northern Hemisphere has declined by approximately 0.3°C, the following question arises: Is it possible that increases in the volume of anthropogenic aerosols could reduce the air temperature at the Earth's surface by several tenths of a degree?

In order to answer that question it is necessary to estimate the influence of changes in the concentration of aerosols in the troposphere on air temperature at the Earth's surface. This is more complex than is the similar influence of changes in the concentration of aerosols in the stratosphere.

The data presented above show that increases in the concentration of anthropogenic aerosols have reduced direct radiation in the Northern Hemisphere by approximately 6 per cent, and have reduced total radiation for average global conditions (for the cloudiness coefficient being equal to 0.5) by 0.5 per cent. Making use of the semi-empirical theory of the atmospheric thermal regime we will find that such a reduction in total radiation produces a decline in average temperature at the Earth's surface by 0.75°C. This corresponds to the case in which the influence of aerosols is limited to the scattering of radiation, and excludes its absorption by aerosol particles. A number of studies have shown that the effect of absorption of short-wave radiation by particles of tropospheric aerosols may compensate to a certain extent for the decline in temperature that is produced as a result of the back scattering of radiation by aerosol particles.

As has already been noted an estimate of the influence of the absorption of radiation by aerosols is made difficult by the absence of reliable data concerning average values of absorption coefficients. Nevertheless it is possible to employ for this purpose the findings of G. Yamamoto and M. Tanaka (1972) who have applied a semi-empirical theory of the thermal regime of the atmosphere in order to calculate reductions in the average global temperature at the Earth's surface as a result of the scattering of radiation by aerosols both in the presence and in the absence of absorption of radiation by aerosol particles. The ratio of changes of temperature in the first case to changes of temperature in the second case (for an average global value of cloudiness with a relatively low coefficient of turbidity) is found to be 0.68. By multiplying that magnitude by the reduction in temperature noted earlier we find that the value of that reduction for realistic conditions is close on 0.5°C, i.e., is comparable to the sum of absolute values of reduction in temperature that has occurred in the 1940s-1960s and its increase attributable to a growing concentration of carbon dioxide. While this calculation is highly approximate it does provide grounds for expecting that changes in the average global air temperature in the course of the 1940s-1960s were largely attributable to man's economic activities.

The influence on current global climatic conditions of all other anthropogenic factors besides atmospheric carbon dioxide and aerosols appears to be relatively small.

Among these factors attention is drawn to the production of energy through various forms of economic activity, which results in an additional heating of the atmosphere and the Earth's surface. The entire energy consumed by man is eventually transformed into heat and the major part of that heat is an additional source of energy for the Earth that contributes to increasing its temperature.

Of all existing components in the current consumption of energy by man only hydropower and the energy contained in firewood and agricultural products represent a transformation of the energy of solar radiation that is absorbed by the Earth each year. Expenditures of energy of this type do not change the Earth's heat balance and do not lead to additional heating.

Nevertheless these types of energy represent a small part of the total energy consumed by man. Other types of energy such as the energy of coal, petroleum, natural gas, and also atomic energy represent a new source of heat that is independent of the transformation of the energy of solar radiation in our present age.

Estimates of the quantity of heat that is generated as a result of man's economic activities have been made in a number of studies by the present author (Budyko, 1962, and so on). This magnitude is not large per unit of the Earth's surface as a whole, and represents approximately 0.01 kcal/cm^2·year. In the case of the more developed industrial regions that magnitude is larger by two orders of magnitude and reaches 1-2 kcal/cm^2·year on territories that cover tens and hundreds of thousands of square kilometers. On the territories of large cities (tens of square kilometers) that value increases by another order or two, i.e., it reaches dozens and hundreds of kcal/cm^2·year.

The influence of this additional production of heat on the Earth's average temperature can be calculated.

It has been noted earlier that one per cent change in the inflow of energy reaching the Earth from the Sun alters the average temperature at its surface by 1.5°C. Assuming that the production of heat as a result of human activities currently represents 0.006 per cent of the total radiation absorbed by the Earth-atmosphere system, we find that the increase in average temperature corresponding to this quantity of heat is approximately 0.01°C. This magnitude is relatively small, yet in view of the considerable unevenness in the location of man-made sources of heat on the Earth's surface in individual regions such an increase in temperature may be considerably larger.

In the absence of atmospheric circulation the temperature of the more developed industrial regions could increase by a magnitude of approximately one degree and in large cities by several tens of degrees, which would clearly make life impossible. The influence of atmospheric circulation reduces the corresponding increases in temperature substantially and that reduction becomes even larger as the area on which the production of additional heat energy is concentrated becomes smaller.

Let us determine how the currently existing system of irrigation influences the average temperature at the Earth's surface. When dry regions are irrigated the albedo at the Earth's surface is reduced by a factor of approximately 0.10. Taking into consideration the relation between the albedo at the Earth's surface and the albedo of the Earth-atmosphere system we find that for low cloudiness such a reduction in albedo at the Earth's surface corresponds to a reduction in the albedo of the Earth-atmosphere system by a factor of 0.07.

The currently irrigated territories constitute approximately 2 million km^2, which represents approximately 0.4 per cent of the overall surface of the Earth. Thus irrigation reduces the Earth's albedo by approximately 0.03 per cent.

It follows from the semi-empirical theory of the thermal regime that change in the Earth's albedo by one per cent alters the average temperature at the Earth's surface by 2.3°C. Accordingly we find that irrigation increases the average temperature at the Earth's surface by approximately 0.07°C. Such a change in temperature may not be significant for the global climate, but it is not negligibly small and as irrigated surfaces increase it may play a definite role.

Aside from irrigation facilities a certain influence on average temperature at the Earth's surface may be attributable to the construction of water reservoirs. The average magnitude of the albedo of the Earth's surface in the presence of water reservoirs in regions possessing plant cover is reduced by approximately the same magnitude as in the case of the irrigation of desert areas.

But since the largest artificial water reservoirs are created in regions with relatively humid climates, in which a more or less significant cloudiness exists, the albedo of the Earth-atmosphere system in such cases changes to a smaller extent than in the case of irrigated regions with low cloudiness. In addition, since the total area of artificial reservoirs is substantially less than that corresponding to irrigated lands, their influence on the average temperature at the Earth's surface is relatively small.

The data in this section show that over the past few decades man's economic activities have begun to influence global climate. It may be presumed that the further development of economic activities will lead to more substantial changes in global climate by comparison with those that have already occurred.

It should be noted that anthropogenic changes in climate are not limited to changes in air temperature. As was noted in Chapter III fluctuations in climate produce substantial changes in the atmospheric precipitation regime, and these changes exert a significant influence on conditions of agricultural production. Since relatively modest changes in the thermal regime may lead to large fluctuations in the moisture regime the problem of anthropogenic changes in climate is already of practical significance in our own time. In the near future this significance will apparently increase, and this makes necessary more detailed studies of this problem.

By influencing the composition of the atmosphere man has begun to influence global processes in the biosphere. At the present time that influence is still not very great, but in the future it may alter natural conditions on our planet very substantially.

X

The Future of the Biosphere

1. The Evolution of the Biosphere

The Biosphere in the Phanerozoic Eon. In order to forecast the influence of human activities on the state of the biosphere it is necessary to rely on numeric models that describe the biosphere's evolution. At the present stage of studies of global ecological processes it is only possible to construct highly simplified models for that complex system that convey knowledge concerning only some of its essential aspects.

Past data concerning changes in the biosphere and its major components play an important part in ensuring the realism of such models. By studying these changes through numeric modelling it is possible to identify cause-and-effect mechanisms in the biosphere's development.

Viewing the biosphere as an integrated ecological system whose state is determined by the circulation of various types of matter, let us consider changes in the oxygen and carbon dioxide cycles for the Phanerozoic eon, i.e., for the interval of time encompassing the last 570 million years.

Paleontological studies show that during the Phanerozoic eon most living organisms received their energy from autotrophic photosynthesizing plants that initially existed only in bodies of water, but during the first half of the Paleozoic era came to be disseminated over continents as well. The activity of these plants depended greatly on the chemical composition and physical state of the atmosphere and of the hydrosphere, which were not constant.

The information concerning climatic changes during the Phanerozoic eon mentioned in Chapter III must be supplemented by data concerning the evolution of the chemical composition of the atmosphere, which was of major importance for the development of living organisms.

As noted in Chapter VI the intensity of photosynthesis depends substantially on the concentration of carbon dioxide in the atmosphere and in water bodies, which has changed considerably in the past. Carbon dioxide enters into the atmosphere and hydrosphere from the lithosphere (in particular from volcanoes) and as a result of a diversity of processes moves into various layers of the Earth's crust.

The difference between the inflow and expenditure of carbon dioxide determines changes in its concentration over the time interval under consideration. A part of the expenditures of carbon dioxide is determined by the accumulation of organic matter in the Earth's crust produced by autotrophic plants. It is in this way that coal and other forms of organic carbon deposits have developed. The quantity of carbon dioxide assimilated in the course of photosynthesis and buried in deposits is proportional to the inflow of free oxygen into the atmosphere and hydrosphere over the corresponding time intervals. The inflow of oxygen that is attributable to a greater value of its inflow by comparison with outflows in the course of activities of living organisms is balanced by expenditures of oxygen for the oxidation of mineral substances.

While in a general way the circulation of carbon dioxide and oxygen during the Phanerozoic eon were more or less similar, the intensity of these cycles changed markedly in accordance with different levels of global productivity of autotrophic plants and different intensities of oxidation of organic and mineral substances.

The creation of organic substances through photosynthesis depended considerably on both the volume of carbon dioxide in the atmosphere and hydrosphere and on the volume of free oxygen. In this connection the biosphere's evolution was largely determined by changes in the volumes of carbon dioxide and of oxygen.

Quantitative studies of these changes meet with major difficulties. One of the first attempts to overcome them is described below. It refers to studies made by the author, A. B. Ronov, and A. L. Yanshin.

Data concerning some of the components of continental sedimentary rock of various ages during the Phanerozoic eon are presented in Table 14. These were derived from data produced by Ronov (1959). All values in that table relate to time intervals of one million years.

The data in Table 14 may be employed to determine changes in the volume of carbon dioxide and oxygen in the atmosphere and hydrosphere.

Since these data do not include data relating to oceans, and since their precision is limited by gaps in geological information, they should be viewed as a relative rather than absolute characterization of the overall intensity of the formation of deposits over different time intervals for the Earth as a whole.

In calculating changes in the quantity of carbon dioxide over time we will take into consideration the fact that expenditures of CO_2 on the formation of sedimentary rock is zero, when that gas is absent in the atmosphere and hydrosphere. Accordingly, when concentrations of carbon dioxide are small, its expenditure increases as concentration increases. The following question arises: Will such a principle also hold for levels of concentration of CO_2 that existed during the Phanerozoic eon? There exists a view that the rate of carbon formation in water bodies increases as the quantity of carbon dioxide in the corresponding waters is reduced. It may be thought, however, that such a relationship cannot hold for average global conditions. For if there existed a positive feedback relation between the concentration of CO_2 and its expenditures for the planet as a whole the volume of CO_2 in the atmosphere and the hydrosphere would change within a wide range. Since the quantity of free carbon dioxide in the biosphere is relatively small, it may be easily seen that increases in expenditures of carbon dioxide in situations in which its volume declines (and declines in its expenditure when its volume increases) would lead to enormous fluctuations in the concentration of CO_2, and this is hardly compatible with a prolonged existence of life on Earth.

It should therefore be recognized that in fact the corresponding feedback relation is a negative one, and this makes it possible to rely on the simple assumption that expenditures of CO_2 are proportional to its current concentration.

Relying on that assumption, and assuming that the data

Table 14

THE COMPOSITION OF DEPOSITS ON CONTINENTS

Stratigraphic interval	Absolute age (millions of years)	Mass in 10^{21} g/million years		
		Vulcanogenic rock	Total quantity of CO_2 in sedimentary rock	Quantity of organic carbon in sedimentary rock
Lower Cambrian	490-570	0.35	0.45	0.009
Middle Cambrian		0.18	0.32	0.005
Upper Cambrian		0.14	0.32	0.003
Ordovician	435-490	0.33	0.27	0.009
Silurian	400-435	0.19	0.22	0.005
Lower Devonian	345-400	0.62	0.24	0.003
Middle Devonian		1.43	0.66	0.018
Upper Devonian		1.76	0.68	0.024
Lower Carboniferous	280-345	1.02	0.70	0.022
Middle and Upper Carboniferous		0.27	0.30	0.010
Lower Permian	235-280	0.95	0.61	0.009
Upper Permian		0.35	0.23	0.005
Lower Triassic	185-235	0.32	0.18	0.004
Middle Triassic		1.10	0.34	0.004
Upper Triassic		0.85	0.28	0.009
Lower Jurassic	132-185	0.33	0.31	0.016
Middle Jurassic		0.38	0.39	0.033
Upper Jurassic		0.31	0.40	0.028
Lower Cretaceous	66-132	0.67	0.32	0.023
Upper Cretaceous		0.65	0.44	0.018
Paleocene	2-66	0.31	0.20	0.012
Eocene		0.37	0.31	0.020
Oligocene		0.17	0.08	0.017
Miocene		0.35	0.18	0.019
Pliocene		0.32	0.07	0.014

in Table 14 concerning the total amount of CO_2 in sedimentary rock are proportional to the corresponding global magnitudes, it is possible to calculate changes in the quantity of carbon dioxide in the atmosphere.

By extrapolating data from that table concerning the quantity of CO_2 in sediments during the Cenozoic era, we find that the corresponding quantity for the past million years is $0.05 \cdot 10^{21}$g. By comparing that value with the current concentration of carbon dioxide in the air, which is equal to 0.03 per cent, we find the corresponding coefficient of proportionality. On that basis we then find the value of the concentration of atmospheric carbon dioxide for all subdivisions of the Phanerozoic eon. These values are presented in Figure 35 in terms of the curve for CO_2.

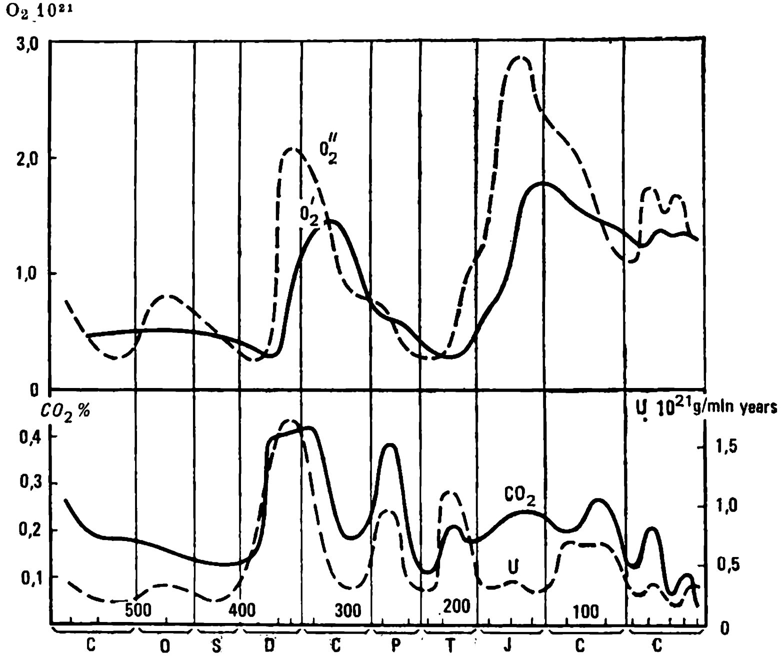

Fig. 35. The Evolution of the Atmosphere During the Phanerozoic Eon (the letters at the bottom refer to geological periods and the Cenozoic era)

This Figure also represents changes in the volume of volcanic rock per unit time interval (the curve *U*). It may be seen that volcanic activity has changed rhythmically and that maximum values were separated by intervals of time of the order of 100 million years. That conclusion accords with the results of theoretical studies of the process of thermal convection in the lithosphere (A. Tikhonov et al., 1969).

These data show that changes in the volume of carbon dioxide are associated with the level of volcanic activity. The highest maximal values of volcanic activity coincide with increases in the volume of carbon dioxide, while reductions in volcanic activity are accompanied by reduction in the quantity of CO_2. It is noteworthy that from the middle of the Cretaceous period the concentration of carbon dioxide has begun to decline, and that process has become more rapid during the Oligocene and especially the Pliocene, at the end of which the volume of carbon dioxide attained its lowest value for the entire Phanerozoic eon. That result is confirmed by the point of view presented in Chapter III as to the influence of declining concentrations of carbon dioxide on reducing temperatures in the Cenozoic era which led to the development of glaciations. It should also be noted that the level of concentration of CO_2 found in these calculations throughout nearly the entire duration of the Phanerozoic eon, which is several tenths of one per cent of the volume of the atmosphere, is optimal from the point of view of the productivity for autotrophic plants. That conclusion confirms the long existing presumption that autotrophic plants have developed in atmospheric conditions in which there was more carbon dioxide than today. It should be noted that the values of concentrations of CO_2 that were found did not attain critical values from the point of view of disrupting vital activities of any living organisms.

These conclusions confirm the realistic nature of the data that have been obtained concerning changes in the volume of carbon dioxide. At the same time the accuracy of these data is limited because of a reliance on a number of approximating assumptions. In particular, one should note the possibility that the rate of reduction of concentrations of CO_2 may be overrated for the Cenozoic era because of the reduction at that time in maritime areas located on territories that today are occuppied by continents.

In calculating changes in the volume of free oxygen it is also possible to rely on the hypothesis that the expenditure of oxygen on the oxidation of mineral substances is proportional to its volume. In that case, however, the magnitudes for the corresponding expenditures are not known, and use should then be made of an equation representing the balance of atmospheric oxygen. This has the following form:

$$\frac{dM}{dt} = A - B, \tag{10.1}$$

where A is the rate of inflow of oxygen resulting from a rate of photosynthesis among autotrophic plants that exceeds losses on the oxidation of organic substances; that rate is proportional to the volume of organic carbon deposited in the Earth's crust; B is the rate of expenditure of oxygen on the oxidation of mineral substances; and $\frac{dM}{dt}$ is the rate of change in the volume of free oxygen over time.

Keeping in mind that for each gram of organic carbon entering into deposits 2.67 g of oxygen enter into the atmosphere, we will assume that $A = \frac{2.67\ C_0}{\alpha}$, where C_0 refers to the data in Table 14 concerning the mass of organic carbon deposited per unit time, α represents a dimensionless coefficient describing the ratio of these values to global values. When data from the Table are close to global ones, the coefficient α is equal to 1, when they are less than the global values, the coefficient α will be less than 1.

In accordance with the hypothesis that has been mentioned earlier we will assume that $B = \beta M$, where β is a coefficient of proportionality.

If the rate of change in the volume of oxygen is small by comparison with the rate of its inflow or outflow (in particular this occurs for small values of the coefficient α) we obtain from (10.1) the formula

$$M = \frac{2.67 C_0}{\alpha\beta}. \tag{10.2}$$

Assuming that during the Pliocene the mass of atmospheric oxygen was close to its current value, and making use of the corresponding value of C_0 it is possible to find the value of the coefficient $\alpha\beta$ and to determine the magnitude of M with the aid of formula (10.2). The result of such a calculation of the magnitude M during the Phanerozoic eon is shown in Figure 35 by the curve O_2''.

Another limiting case of calculating the change in the mass of oxygen over time refers to the condition $\alpha=1$.

Assuming that at the initial moment of time $M=M_0$ we find from (10.1) that

$$\frac{2.67C_0}{\beta}-M=\left(\frac{2.67C_0}{\beta}-M_0\right)e^{-\beta t}. \qquad (10.3)$$

By making use of the condition of equality between the volume of oxygen at the end of the Cenozoic era and its current value in determining the value of the coefficient β we apply this equation to data concerning the quantity of organic carbon in order to calculate changes in the volume of oxygen during the Phanerozoic eon. In making such a calculation it is appropriate to assume a particular value for the volume of oxygen at the beginning of the first time interval. After that it is possible to calculate changes in that volume during that period. The value of the mass of oxygen obtained through corresponding calculations for the end of the interval will then serve as the initial value for the subsequent time interval. In this way all values for the parameter M_0 needed for the calculations are determined in the course of the calculation itself, except for the initial value which relates to the beginning of the Phanerozoic eon.

It may be shown that the choice of the magnitude for that initial value does not influence substantially subsequent values of the calculations of changes in the volume of oxygen. Thus, for example, if one admits two extreme assumptions concerning the magnitude of the volume of oxygen at the beginning of the Cambrian, namely that it is either equal to zero, or else to its current value, the result of the calculation for subsequent periods will practically coincide following the Ordovician. Keeping this in mind we will assume that during the Cambrian period the magnitude $\frac{dM}{dt}$ was substantially smaller than the inflow of oxygen into the biological cycle, and in this connection we will employ formula (10.2) in order to calculate the value of M_0 relating to average conditions during the Cambrian period. Subsequently we will calculate changes in the mass of oxygen according to formula (10.3) for each time interval, until the end of the Phanerozoic eon.

The results of such a calculation are presented in Figure 35 by the curve O_2'. A comparison of the curves O_2', and O_2''

shows that they are relatively similar, and that in the first case the amplitude of changes in the volume of oxygen is reduced and the position of maximum values and minimum values lags somewhat by comparison with the second case. It may be presumed that actual changes in the volume of oxygen occur within the interval between these two limiting curves.

While it follows from these findings that changes in the quantity of oxygen during the Phanerozoic eon were not as considerable as was supposed in earlier studies, these changes must have exerted a perceptible influence on the conditions of the existence of living organisms.

It may be seen from Figure 35 that a large volume of oxygen was contained in the atmosphere during the Upper Devonian and the Carboniferous period and also during the Upper Jurassic and the Cretaceous period. It is possible that this explains the wide dissemination of insects during the Carboniferous period (including gigantic forms), for which a high level of metabolism is characteristic at times of activity, and also the dissemination of many forms of gigantic reptiles during the second half of the Mesozoic era. The greatest reductions in the volume of oxygen occurred during the Triassic period when substantial changes in the fauna of land vertebrates occurred and a number of large forms of amphibians and reptiles became extinct.

Because of the limited reliability of initial data and also of the use of approximating assumptions, the calculated values shown in Figure 35 contain substantial errors. It may be thought, however, that they convey correctly the principal features of the atmosphere's evolution during the Phanerozoic eon. Let us note the following regularities governing that evolution.

1. The highest maximal values for volcanic activity coincide with increases in the volume of carbon dioxide, while reductions in volcanic activity during the Upper Cretaceous and in the Tertiary period are accompanied by reductions in the volume of CO_2. This confirms the substantial influence of volcanic activity on the balance of carbon dioxide.

2. Throughout the major part of the Phanerozoic eon the concentration of carbon dioxide in the atmosphere was approximately 0.1-0.4 per cent. Since for concentrations of this magnitude the productivity of most autotrophic plants ap-

proaches a maximum value it may be assumed that this property of plants emerged as a result of their evolutionary adaptation to those environmental conditions that existed during the age when they were formed.

3. Towards the end of the Mesozoic era concentrations of carbon dioxide began to decline, and that process became more rapid during the Oligocene, and especially the Pliocene, at the end of which the volume of carbon dioxide attained its lowest value for the entire Phanerozoic eon. This reduction in concentration of carbon dioxide was probably one of the major causes for the cooling that led, in particular, to the development of Quaternary glaciations.

4. Increases in the volume of oxygen during the Phanerozoic eon were not even. Calculated data show that in a number of cases increases in volcanic activities were accompanied by reductions in the volume of oxygen.

The dependence of the overall mass of living organisms on concentrations of carbon dioxide should be noted. The major part of biomass consists of substances produced by autotrophic plants whose quantity is regulated by their photosynthesis.

It follows from the data in Chapter V that such a dependence exists in cases in which the water regime, the availability of minerals and other factors do not constrain photosynthesis appreciably. It is clear, nevertheless, that even in the presence of additional factors limiting photosynthesis there is a definite relationship between the mass of plants and the quantity of carbon dioxide, which is particularly evident for small changes in concentration of CO_2.

It follows from this that in those geological periods in which the concentration of carbon dioxide in the air was substantially larger than it is today the mass of plant cover on land could be larger than the present one. This refers above all to the plants of sufficiently humid zones. The total volume of biomass of autotrophic plants in the oceans in ages characterized by high content of carbon dioxide could also have been larger than it is at present.

Thus the process that led to a reduction in the volume of carbon dioxide during the Cretaceous and the Cenozoic appears to have been accompanied by reduction in the mass of autotrophic plants and therefore of the entire mass of living organisms on our planet.

Prospects for a Natural Evolution of the Biosphere. One of the factors that will determine the state of the biosphere in the future is a possible change in climate. Let us pause to consider how climatic conditions may change in the absence of influences of the part of man.

Studies concerned with climatic changes have shown that these are especially responsive to the following causes:

1. Astronomical factors. Changes in the position of the Earth's surface in relation to the Sun produce fluctuations in climate with time periods of thousands and tens of thousands of years.
2. The composition of the atmosphere. In the Tertiary and the Quaternary period a definite influence on the climate was exerted by a reduction in the contents of carbon dioxide in the atmosphere. In view of the rate at which that reduction and the corresponding changes in air temperature took place it may be concluded that the influence of natural changes in carbon dioxide content on the climate was substantial for time intervals greater than one hundred thousand years.
3. The structure of the Earth's surface. Changes in relief and associated changes in the position of maritime shores may alter climatic conditions substantially over large areas for periods of time ranging from hundreds of thousands to millions of years.
4. The solar constant. Leaving aside the questions concerning the existence of short-periodic fluctuations in the solar constant that influence the climate, one should recognize the possibility of slow changes in solar radiation produced by the Sun's evolution. Such changes may influence climatic conditions substantially over periods of the order of hundreds of millions and billions of years.

Among other external factors influencing the climate one should mention volcanic activity whose fluctuations may change climatic conditions both over short time intervals (of the order of several years) and for much longer time periods.

Together with changes attributable to external factors climatic conditions change as a result of autonomous fluctuations in processes within the atmosphere-ocean-polar ice system. Such changes refer to periods of the order of years and decades and possibly also of hundreds and possibly thousands of years.

Changes in the position of the Earth's surface in relation to the Sun may be calculated with considerable accuracy.

Such calculations (Sh. Sharaf and N. Budnikova, 1969; A. Vernekar, 1972) make it possible to estimate fluctuations in the inflow of solar radiation during warm seasons from which the position of the boundaries of polar ice depends most. As has been noted by M. Milankovich (1930) ice boundary shifts are especially sensitive to fluctuations in radiation at latitudes close to polar circles.

It follows from the data produced by these calculations that in 10,000-15,000 years at "critical latitudes" of the Northern Hemisphere there will occur a perceptible reduction in radiation whose magnitude will be approximately two-thirds of the reduction that took place during the last Würmian glaciation. Such a reduction in radiation may lead to the development of a new ice age marked by continental glaciers covering a smaller area than during the Würmian period.

Subsequently, a perceptible increase in radiation will occur that should lead to a retreat of ice sheets, and then a reduction in radiation at critical latitudes will again take place approximately 50,000 and 90,000 years from now. The amplitude of these reductions in radiation will increase, approaching the magnitude of the reduced radiation that caused the last Würmian glaciation.

Much larger reductions in radiation at critical latitudes in the Northern Hemisphere will take place in several hundred thousand years. This points to the possibility of enormous glaciations in these times.

The influence of astronomic factors on the future climate may be greatly intensified as a result of a further reduction in the carbon dioxide content in the atmosphere. Since over many millions of years there has prevailed a tendency towards a reduction in the concentration of carbon dioxide in the air, it is probable that in the absence of effects on climate originating in economic activities the volume of carbon dioxide in the future will continue to decline. The rate at which this decline will occur may be estimated on the basis of data relating to the reduction of the concentration of carbon dioxide in the Tertiary period.

Applying such an extrapolation it is possible to estimate over what time interval the concentration of carbon dioxide in the atmosphere could decline from the value that existed before the age of rapid industrial development, equal to

0.029 per cent, to a magnitude at which a full glaciation of our planet is possible, namely 0.015 per cent.

Data relating to changes in the volume of CO_2 in the Tertiary period indicate that this time period will be approximately 10 million years, but data relating to the end of the Tertiary period (Miocene-Pliocene) suggest that it may be much shorter and perhaps will not exceed one million years.

If we also consider that, aside from a general tendency to a development of glaciations, attributable to reductions in the concentration of carbon dioxide, there exist periodic increases in ice sheets, attributable to changes in the position of the Earth's surface in relation to the Sun, then that estimate should be reduced. The probable time for the emergence of a full glaciation of the Earth with due account of this particular effect will be several hundred thousand years.

In the light of these estimates it is possible to conceive the following hypothetical image of the possible evolution of the biosphere in the future.

The continuing decline in the concentration of carbon dioxide will be accompanied by gradual reductions in the productivity of autotrophic plants and a reduction in the overall mass of living organisms on Earth. At the same time there will be a gradual widening of the zone of polar glaciations which move towards lower latitudes as ices ages begin.

The productivity of autotrophic plants and the general volume of biomass on Earth will be reduced by not more than one-half when the ice cover will reach a critical latitude before spreading to the Equator as a consequence of autonomous development. As a result, a full glaciation of the planet will take place that will be attributable to low negative temperatures at all latitudes.

A full glaciation may lead to a cessation of all biological processes on our planet. This presumption is based on the fact that during the prolonged period of time that the Antarctic glacier has existed no living organisms have developed that could live continuously in the central regions of that glacier. Following a spread over a relatively short period of time of the climatic conditions of Central Antarctica to the entire globe it would be difficult to expect that living organisms would develop that could adapt themselves to such unfavourable conditions.

A full glaciation of our planet corresponding to the thermal regime denoted in Figure 13 by the point B' should be extremely stable, since air temperature at all latitudes will be substantially lower than the freezing point of water. At the same time it is possible that in the distant future such a glaciation may vanish as a result of a gradual increase in the solar constant. The authors of a number of astronomical studies have stated the presumption that in the course of the Sun's evolution the temperature at its surface and the quantity of radiated energy will increase. That effect may attain a perceptible value in several billion years (M. Schwarzschield, 1958; A. Ångström, 1965; O. White, 1967).

It may be seen from Figures 12 and 13 that the destruction of planetary glaciation requires an increase in the solar constant by approximately 40 per cent. In such a case ice will melt at all latitudes and air temperature will increase by more than 100°C, attaining an average value greater than 50°C. Since for further increases in the solar constant the temperature of the Earth's surface will continue to increase, the melting of ice is hardly likely to produce conditions that are favourable for a repeated emergence of life on Earth.

Keeping in mind that the question relating to the Sun's evolution has not yet been fully clarified, the hypothetical character of such changes in the thermal regime must be noted. It seems interesting, however, that should such a thermal regime take place it will pass through stages corresponding to a large part of the hysteresis curve in Figure 13, ranging from the ice-free regimes of the Mesozoic era (slightly above point E) to contemporary conditions (point 3), conditions of full glaciation (from point B to point A) followed by the destruction of ice (point A').

For a number of reasons such an estimate of the time corresponding to full glaciation of our planet is very approximate. In addition to the simplified nature of the model of the atmospheric thermal regime that was employed in arriving at such an estimate, the accuracy of the corresponding calculations is substantially limited by the application of methods for determining the rate of decline of carbon dioxide in the atmosphere which are rather conventional.

It may be thought that these findings are primarily of interest as an illustration of the assumptions stated earlier concerning the specific nature of Quaternary glaciations

(Budyko, 1971). It is possible that unlike the glaciation of the Permocarboniferous and other ancient glaciations, Quaternary glaciations were not temporary episodes in the Earth's evolution, but rather the beginning of a transition from a stable ice-free climatic regime to a still more stable regime of a full glaciation of the planet. The duration of such a transitional period, which could bring to an end the existence of a biosphere, is quite small by comparison with the overall duration of existence of life on our planet.

The probability of the development of a full glaciation of the planet in the course of the Earth's natural climatic evolution increases if one considers the existing tendencies towards an increase in the areas of continents at high and middle latitudes in the Northern Hemisphere.

By making use of the paleogeographic maps of V. N. Saks (1960) and of other authors it is possible to establish that towards the end of the Jurassic period oceans covered nearly one-half of the latitudinal circle at 70° North latitude, and this made possible a free circulation of sea waters between moderate and polar latitudes. In the beginning of the Cretaceous the area occupied by continents began to increase gradually at that latitude and this continues to take place at the present time. As a result, today continents account for $^5/_6$ of this particular latitudinal circle. Thus the width of straights linking seas at middle and high latitudes of the Northern Hemisphere has declined to one-third of its original value. A corresponding substantial reduction in the meridional transfer of heat into polar latitudes by ocean currents has taken place and conditions have emerged for the development of Arctic glaciation.

There are no grounds for expecting that the tendency towards a growing isolation of the Arctic basin which has already existed for many millions of years will cease in the near future. A continuation of that tendency means that in the future the probability of new glaciation processes and the possibility that they may reach a critical latitude will increase.

It should be noted that the effect of changes in the Earth's surface structure on future climate will probably express itself after the effect of a reduction in the carbon dioxide content in the atmosphere and may therefore prove to be unimportant. Yet in view of the approximate nature of many of the corresponding estimates it is impossible to say that the relative significance of these two factors in future climatic changes has been established.

Figure 36 represents a hypothetical view of changes in average global air temperature at the Earth's surface in the future which reflects the considerations that have just been discussed. It also represents changes in average global temperature in the past. For purposes of exposition the scale of the time axis on that diagram has been magnified by one thousand times for the past million years and the next million years. The interval for changes in average temperature within which the most favourable conditions for the development of biological processes exist is represented by the shaded area.

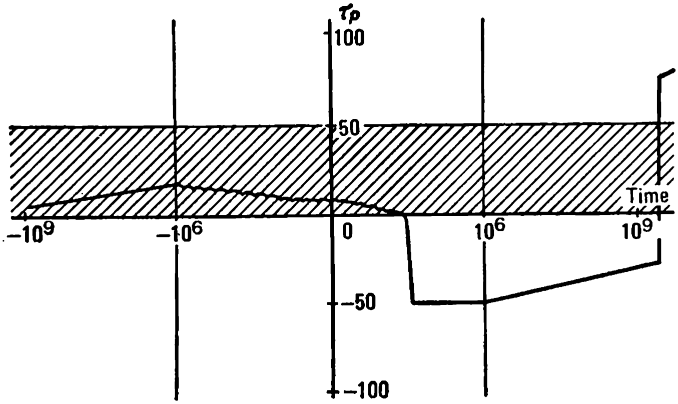

Fig. 36. Change in the Average Global Air Temperature

The limitations that attach to the diagram in connection with the underlying hypotheses should be emphasized. Nevertheless the conclusion that follows from the data that have been considered concerning the low stability of the biosphere in relation to fluctuations in the external factors that influence it deserves attention. Relatively modest changes in these factors may bring about a destruction of life on Earth and accordingly a destruction of the biosphere. It would seem that such a conclusion does not accord well with the actual prolonged existence of the biosphere over three billion years.

In earlier studies (Budyko, 1971) the possibility was noted that the prolonged existence of the biosphere in the past might be attributable to a coincidence of factors that are favourable to the existence of life, which is rare statistically, even though these factors have remained active over a long period of time. In such a case the probability of finding ancient biospheres that include higher forms of life in regions of the Universe that are close to the solar system would be extremely small. This probably explains the fact that all attempts to obtain information about extra-terrestrial civilizations have so far ended in failure.

The development of mankind, which took place extremely rapidly from a geological point of view, has altered fundamentally the prospects for the further existence of the Earth's biosphere.

Suffice it to note that during the last several decades alone the burning of various types of fuel has increased the concentration of carbon dioxide in the atmosphere by 0.003 per cent. Such an increase in the quantity of carbon dioxide compensates for a reduction that has taken place over more than 200,000 years. Thus the activity of man has altered the direction of the process of changes in concentration of carbon dioxide in the atmosphere and has increased the rate at which that process takes place by many thousand times. Even though in the present case man's influence on the climate was inadvertent, it has already played an important role in forestalling the further development of glaciations.

Let us consider the not very probable case in which man's influence on the atmosphere will cease in the future. It may be assumed that under such conditions the increase in concentration of carbon dioxide in the atmosphere that has taken place in the past century will postpone the glaciation of the planet by several thousand years. It is evident that should the present scale of influence on the atmosphere continue or increase, the possibility of a global glaciation may be excluded.

At the same time the rapidly increasing influence of economic activities on the climate makes possible perceptible changes in the global climate in the relatively near future. Such changes, of course, cannot have the type of catastrophic consequences that a full glaciation of the planet might produce, yet they may produce a number of major difficulties for man's economic activities.

In this connection the nature of future anthropogenic changes in climate deserves considerable attention.

2. *The Biosphere in the Near Future*

Anthropogenic Changes in Climate. The problem relating to the distant future of the biosphere that was considered in the preceding section is largely a matter of general interest. The greatest practical interest attaches to the problem of studying possible changes in natural conditions in the next several decades, i.e., for the period that should be considered in the long-term economic development plans now being drawn up.

It has already been noted that man's activity has already begun to influence large-scale atmospheric processes, and that this influence is increasing rapidly. The need to identify the nature of the consequences that may follow from that process in the next 50-100 years is evident.

Let us examine the influence of three basic factors on the climate of the next century:

1) A growth in the production of energy consumed by man;

2) An increase in the content of carbon dioxide in the atmosphere;

3) Changes in the concentration of atmospheric aerosols.

It was noted in one of our studies (Budyko, 1962) that increases in the production of energy ranging from 4 to 10 per cent per year will lead to a situation in which in 100-200 years the quantity of heat produced by man will be comparable to the magnitude of the radiation balance on the surface of all continents. Clearly in such a case enormous changes in climate will take place.

Some idea about the possible influence of human activities on the climate of the future may be derived from the data presented in Figure 37. In that Figure curve *1* represents the secular course of deviation from normal of average air temperature in the Northern Hemisphere, based on observations. Curve *2* represents the results of calculations of changes in the planet's average temperature for a rate of increase in the output of energy of 6 per cent per year. That calculation is based on the utilization of a semi-empirical model of the thermal regime of the atmosphere with the assumption that changes in temperature occur as a result of changes in the inflow of heat, while all other factors influencing the climate remain constant, including the albedo of the Earth-atmosphere system.

It may be seen from Figure 37 that the temperature increase attributable to a growth in the production of energy will become greater in the first half of the twenty-first century than changes in temperature that had occurred in the first half of the twentieth century as a result of natural factors. Subsequently increases in temperature will grow rapidly, and this will produce major changes in global climate. Thus the growth in production of energy consumed by man may exert a substantial influence on the climate of the future.

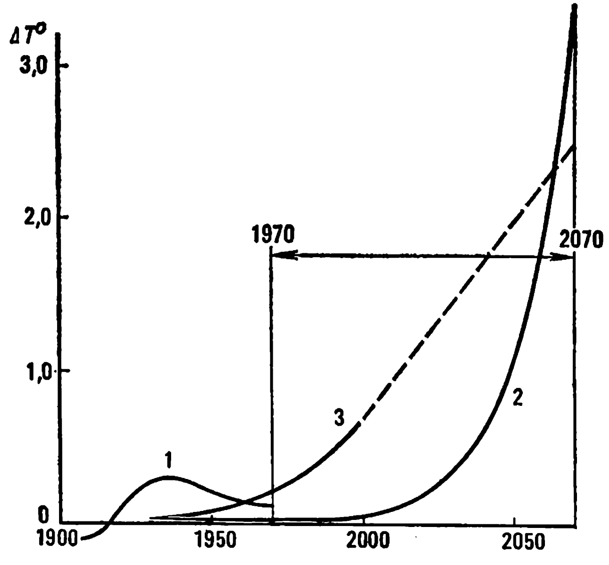

Fig. 37. Changes in Air Temperature Anomalies at the Earth's Surface

It has already been noted that the second factor that may change climate perceptibly is the growth in the concentration of carbon dioxide in the atmosphere. Making use of the numerical model representing the balance of carbon dioxide Machta and other authors have concluded that by the year 2000 the concentration of carbon dioxide in the atmosphere will increase by approximately 15-20 per cent (*Inadvertent Climate Modification*, 1971).

The corresponding calculations show that under the influence of such a factor the average air temperature at the Earth's surface may increase by approximately 0.5°C by the year 2000, by comparison with 1970, while the average temperature at high latitudes may increase by approximately 1°C. Changes in global temperature corresponding to the results of such a calculation are represented in Figure 37 by the solid part of curve *3*.

It may be seen from that Figure that by the year 2000 the changes in temperature attributable to a growth in the concentration of carbon dioxide may exceed the anomalies attributable to natural fluctuations in global temperature that were observable during the first half of the twentieth century. It may be expected that a continuation in the growth of carbon dioxide concentration in the twenty-first century will lead to still greater changes in the climate.

Among other factors that are associated with economic activities and that may bring about fluctuations in climate one should note changes in the concentration of aerosols in the atmosphere. Data represented in Chapter IX show that increases in the atmospheric aerosol content attributable to economic activities exert a perceptible influence on today's climate. It is evident that a further growth in the concentration of aerosols may lead to a substantial change in climate, but it is difficult to forecast the quantity of atmospheric aerosols in the future.

In recent years major efforts have been undertaken in a number of countries to create a system for protecting the atmosphere from pollution, to check the growth of concentration of anthropogenic aerosols so that its value towards the end of the century will either remain at the present level or may even decline. It may be seen from Figure 34 that from the beginning of the 1960s a reduction in the transparency of the atmosphere has ceased, and this may indicate that the volume of anthropogenic aerosols has become stable.

An estimate has been given in Chapter IX of the influence of anthropogenic aerosols on average global temperature, and the conclusion was made that their contemporary volume decreases the average temperature at the Earth's surface by approximately 0.5°C. To a significant extent that magnitude is compensated by increases in temperature attributable to larger volumes of carbon dioxide.

Passing on to possible changes in climate in the next century one should above all discuss the consequences of a further growth in the concentration of carbon dioxide in the atmosphere and of increased energy production.

It may be seen from Figure 37 that for a yearly growth in the production of energy by 6 per cent a rapid increase in average global temperature will begin in the middle of the twenty-first century. This may produce great difficulties for economic activity. Thus an uncontrolled growth in the production of heat will lead to the appearance of a specific type of "heat barrier" on the road to further development of energy production.

It should be noted that the assumptions embodied in the construction of curve *2* on Figure 37 cannot be viewed as unrealistic. An increase in the production of energy by 6 per cent per year corresponds to the rate of increase in that output that has taken place in recent years (*Inadvertent Climate Modification*, 1971). The value for the additional inflow of heat per unit area in 2070 obtained in that calculation is relatively close to the quantity of heat that is currently contributed by individual regions within the industrially more developed countries.

As a result, the possibility that a heat barrier will be reached within a period of approximately 100 years may be considered as one of the major problems that will confront technology and energy production in the relatively near future.

The question relating to ways in which that problem may be overcome calls for specialized discussions that lie beyond the framework of the present study. In the present context one may only recall the considerations mentioned by N. N. Semyonov concerning the desirability of a wide use of solar energy in the future for economic purposes, since this will make it possible to limit the excessive heating of the Earth's atmosphere.

Long before such a heat barrier is reached a major influence on natural conditions may be exerted by increases in temperature attributable to the growth of carbon dioxide concentration. The conviction that such a heating is inevitable has been expressed several years ago (Budyko, 1972) and has been confirmed in very recent times (Budyko, Vinnikov, 1976), when it was found that since the end of the 1960s an increase has been recorded in the average air temperature in the Northern Hemisphere. This increase probably explains the increase in the frequency of droughts in a number of countries at middle latitudes.

In studying changes in climate attributable to increases in global temperature the question relating to the state of the polar ice covers is of major importance.

In estimating changes in the area covered by polar ice use has been made of the data presented in Figure 37 and of the semi-empirical theory of the atmospheric thermal regime. This has made it possible to calculate the future positions of the average boundary of Arctic sea ice. The outcome of such calculations is presented in Figure 38, in which the vertical axis represents the average latitude of the boundary of sea ice in the Northern Hemisphere (Budyko, 1972).

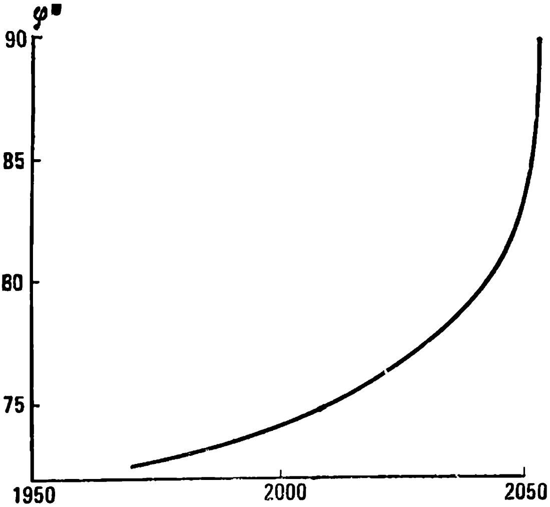

Fig. 38. Changes in the Boundary of Sea Polar Ice

It may be seen from that graph that by the year 2000 the average boundary of polar ice will recede to the north by approximately two degrees. Calculations show that for such a change in the corresponding boundaries a further increase in air temperature will occur at high latitudes, particularly during cold seasons. That increase in temperature will be appreciably greater than the increases in temperature observable in the 1920s and 1930s. It should be noted that since the publication of the corresponding study it was found that a reduction in ice covers in the Arctic has in fact taken place that was associated with climatic changes.

It may be seen from Figure 38 that for the corresponding assumptions the time that is required for a full melting of sea ice in the Arctic is approximately 75 years.

Several indicators of changes in climatic conditions associated with the melting of polar ice are presented in Figure 39. That figure shows changes in temperature at middle and high latitudes of the Northern Hemisphere by the year 2050, i.e., for the time of the completion of melting of polar ice, that were calculated with the help of the formulae of the semi-empirical theory of thermal regime. It may be seen from this diagram that a full melting of polar ice will lead to very large changes in the thermal regime at high latitudes.

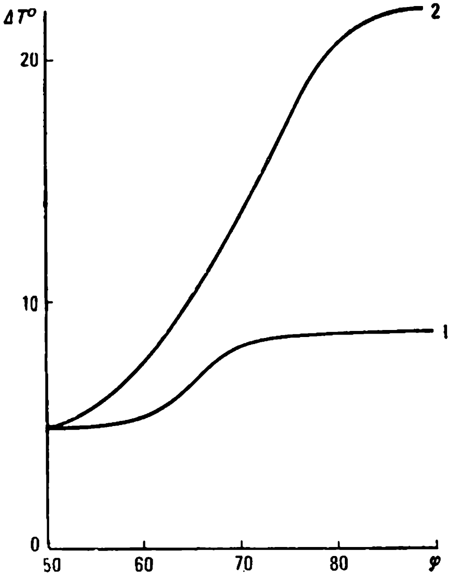

Fig. 39. Changes in Average Temperature at Various Latitudes Following the Melting of Polar Ice (1 — warm season half year; 2—cold season half year; the horizontal axis represents degrees of Northern latitude)

A comparison of the data in Figure 39 with paleoclimatic data shows that in the absence of ice in the Arctic the thermal conditions at middle and high latitudes of the Northern Hemisphere will be similar in many respects to the conditions that existed towards the end of the Tertiary period millions of years ago. In that connection a transition to new thermal conditions may be viewed as a return to the climatic regime of the Prequaternary period. The very short time during which a return to earlier climatic conditions may take place should be noted—it is approximately 100,000 times more rapid than is the natural process of cooling that has prevailed in recent geological periods.

It is apparent that changes in the climate associated with the melting of sea polar ice will not be limited to increases in air temperature. One of the possible consequences of the melting of sea ice may be gradual destruction of the glaciers in Greenland and the Antarctic. This would bring about an increase in the level of the ocean. Similarly, changes in the water exchange on continents may also have important practical consequences.

It was shown in Chapter III that even if a relatively small part of the Arctic ice cover was melted during the 1920s and 1930s the water exchange on continents was greatly influenced at middle latitudes of the Northern Hemisphere. Corresponding changes have produced an increase in the frequency of droughts in a number of regions and a change in river run-off and levels of inland seas.

Many aspects of changes in climate during the first half of the twentieth century were of considerable practical significance. In this connection it may be thought that changes in climate that had begun in recent years would exert a major influence on various aspects of economic activity in a number of countries. A further intensification of climatic changes associated with the melting of polar ice could acquire the dimensions of a climatic catastrophe unless special measures are introduced in adapting economic activities to these changes.

The considerations bearing on climate conditions in the future that were presented should be supplemented by an estimate of the influence of expected changes in the quantity of anthropogenic aerosols.

Although it has already been observed that it is difficult to forecast these changes it may, however, be assumed that a large increase in the quantity of anthropogenic aerosols in the future is not very probable since mankind has already become conscious of the fact that it might produce extremely unfavourable conditions for life itself in regions in which major sources of such aerosols exist, namely in cities and industrial centers. Accordingly, it is hardly possible to think that towards the end of the twentieth century and especially in the twenty-first century a growth in the volume of anthropogenic aerosols will compensate the trend towards increased air temperature attributable to the growing production of energy and to increases in the concentration of carbon dioxide in the atmosphere.

Let us note that aside from anthropogenic changes in the heat and water balances the possibility exists that there may be changes attributable to economic activity in the quantity of oxygen in the atmosphere which is expended on the burning of various types of fuel and other productive processes (Davitaya, 1971).

Calculations show that even before a perceptible reduction in the quantity of oxygen is reached very major changes in climate will take place as a result of the growing concentration of carbon dioxide and increased output of energy representing an additional source of heat for the atmosphere. Thus the problem of preserving the atmospheric oxygen may emerge only at later stages in the ecological crisis that is being produced by human activities.

In discussing the data which have been cited and which characterize possible changes in climate during the next century it is important to keep in mind that in many cases they are based on hypotheses.

Above all it is important to repeat the point made earlier that the use in such calculations of an extrapolation of existing rates of growth in the production of energy and concentration of carbon dioxide in the atmosphere is a possible but not necessarily correct hypothesis in relation to future economic activities. Moreover, it may be expected that should further studies confirm the conclusion that sharp changes in climate are possible during the next century, measures will then be taken to limit the influence of industrial development on climatic conditions.

The Productivity of Agriculture. Among the many possible consequences of anthropogenic changes in climate their influence on agriculture is of the greatest practical significance. It is known that the world's rapidly growing population is poorly supplied with food. As a result of a low level of development of agricultural production hundreds of millions of people in many countries of Africa, Asia, and America continue to suffer from hunger, and growth in the production of food in these countries lags behind population growth.

The major components of today's global food balance include wheat (the major regions of wheat production are concentrated at middle latitudes of Europe, Asia, and North America) and rice (which is grown in a number of subtropical and tropical areas). The overall territory engaged in agricultural production is approximately 10 per cent of the surface of continents. Only a part of that territory yields the high harvests that are made possible by mechanized agricultural practices, highly productive strains and a sufficient quantity of fertilizers.

In recent years food prices have increased rapidly on the world market in connection with the reduction of available reserves of grain. This reflects a worsening of the food problem in many countries. Under such conditions even relatively modest fluctuations in those aspects of climate that influence agricultural production may lead to major economic consequences.

A major influence on fluctuations in harvests is exerted by changes in precipitation regimes. Thus, for example, the reduction in precipitation during the period of increased temperatures in the 1930s has reduced grain harvests in the United States by 10-15 per cent (average values for five-year intervals). Changes in climate in the direction of cooling, such as the one that took place in the 1940s-1960s, had a favourable effect on the precipitation regimes in the hinterlands at middle latitudes.

Increases in temperature differences between the Equator and polar regions resulting from cooling processes have increased the intensity of atmospheric circulation and intensified the water exchange between oceans and the hinterlands regions. While such changes in climate do not preclude major droughts, they greatly reduce their recurrence by comparison with periods of increasing temperatures.

There is no doubt that the major part of the very large increase in harvests that has been achieved in many countries at middle latitudes during recent decades is attributable to progress in agrotechnics. But it is possible that a certain part of that increase is also attributable to a growth in the volume of precipitation. New increases in temperature may exert a perceptible influence on agricultural productivity.

During the past fifteen years global food production has increased by an average of two per cent per year. This value is comparable to the yearly increment in population and this has made it possible to maintain a certain balance between the production and consumption of food (even though this did not preclude intense food shortages in many developing countries).

Should changes in climatic conditions reduce the total production of food by only one per cent per year this could lead to the death from famine of tens of millions of persons who already experience an intense need for food products. Statistical data show that during the past fifteen years unstable weather conditions have occasionally reduced world food production by more than one per cent. Shortages of food in years of poor harvest are partly met from grain reserves, but such a method for regulating the food balance can obviously operate only in the case of short-term fluctuations in conditions of agricultural production.

In order to overcome unfavourable effects of long-term changes in climate on harvests it is necessary to intensify progress in agricultural technics still further or else apply methods for influencing the climate. Since climate in fact influences all large-scale processes in the biosphere that second alternative in fact constitutes a regulation of the biosphere. It may be expected that science and technology will provide a solution to that problem in the relatively near future.

The Problem of Regulating the Biosphere. In order to forestall unfavourable climatic changes appearing as a result of economic activity a number of measures are being carried out among which a major role is played by those directed against atmospheric pollution.

It should be noted that as a result of various measures adopted in many countries, including the cleaning of air used by industrial enterprises, transport, and heating facilities, a reduction in the level of air pollution has been achieved in a number of cities in recent years, even though in many other regions air pollution continues to increase. This points to the great difficulties that attach to preventing a growth in the quantity of anthropogenic aerosols in the atmosphere.

Activities to prevent increases in the content of carbon dioxide in the atmosphere and the growth of heat contributed by transformations of energy consumed by man would be more difficult and have not yet been considered. There do not exist simple technological means for solving these problems, especially the second one, aside from limiting the consumption of fuel and of most types of energy. In the next few decades, however, this appears to be incompatible with further technological progress.

In order to preserve existing climatic conditions it will therefore be necessary to regulate climate in the near future. It is clear that the regulation of climate may also be employed in order to prevent unfavourable natural climatic fluctuations, and that in the future it may be employed to change global climate in directions intended to serve the interests of man.

Possible ways in which climate may be influenced have been discussed in a number of studies containing a variety of proposals.

Ge. K. Fyodorov has pointed out on several occasions (1958, 1972, etc.) that the most promising approaches relate to influencing large-scale atmospheric processes possessing a low stability, that is, a high sensitivity to relatively small changes in external factors. That principle has yielded positive results in several cases in influencing cloudiness, precipitation, fog, and other weather phenomena.

Fyodorov has cited an interesting example of the manner in which the dispersion of clouds has altered perceptibly the radiation balance over a relatively large territory. Such experiments show that prolonged influences on the weather may change climatic conditions. This has led him to conclude that in principle it is possible to influence the climate of the entire globe or else of major parts of the globe.

There exist a number of studies in which various projects for influencing the climate are considered. Usually they relate to changes in local climates. The outcomes of their application are often not clear and the methods for carrying them out are insufficiently developed.

While recognizing that the problem of regulating climate is still at an early stage of development let us note that in recent years studies have been published that are concerned with the problem of influencing global climate through changes in the mass of aerosols in the stratosphere (Budyko, 1974; Budyko et al., 1974).

In recent years it was established that there exists a layer of aerosols in the lower stratosphere with particles whose average size is of the order of tenths of a micron and that are primarily composed of sulphur compounds.

An estimate of the influence of that layer on the flow of the short-wave radiation reaching the Earth's surface was presented in Chapter IX. It follows from that estimate that relatively modest fluctuations in the volume of aerosols in the lower stratosphere alter the "meteorological solar constant" perceptibly and, hence, the average air temperature at the Earth's surface.

Let us consider the possibility of a change in climatic conditions through an increase in the concentration of aerosol particles in the lower stratosphere. In that connection let us calculate the mass of aerosols whose introduction into the stratosphere could produce changes in climate on a scale comparable to the one that occurred as the result of increased temperatures in the 1920s and the 1930s, but having an opposite direction, i.e., producing a decline in the average temperature at the Earth's surface.

It may be seen from observational data that this would require a reduction in the average value of direct radiation by two per cent, which represents a reduction in the total radiation of 0.3 per cent. Assuming that the average magnitude of the mass of aerosols that reduces total radiation by one per cent is $0.8 \cdot 10^{-6}$ g/cm^2, we find that the volume of aerosols in the Northern Hemisphere that is required for such an objective must be increased by 600,000 tonnes.

Different technical methods may produce such an increase in the volume of aerosols. Let us consider one of them, which assumes that the aerosol particles in the stratosphere that influence the magnitude of "meteorological solar constant" are sulphates.

In order to increase the concentration of sulphate particles it is sufficient to introduce a different quantity of sulphur dioxide at the level of aerosol layer, which is subsequently transformed into drops of sulphuric acid. In order to create 600,000 tonnes of sulphuric acid approximately 400,000 tonnes of sulphur dioxide are needed. This may be produced as a result of burning 200,000 tonnes of sulphur.

Let us assume that the average life-time of aerosol particles in the atmosphere is two years. In order to maintain the needed concentration of aerosol particles consisting of pure sulphuric acid it is then necessary to deliver 100,000 tonnes of sulphur into the lower stratosphere each year. Because of the hygroscopic nature of concentrated sulphuric acid, however, its drops will absorb water vapour in the air and increase their mass, while reducing their specific weight. As a result of that effect, the quantity of reagent needed to bring about the indicated reduction in the total value of radiation declines substantially. Since the exact magnitude of the concentration of sulphuric acid in drops of stratospheric aerosols is not known, we will provisionally assume that it is 70 per cent. In such a case approximately 60,000 tonnes of sulphur per year are needed to produce the required change. Let us note that because of the reduction of the specific weight of aerosols as sulphuric acid is diluted with water this magnitude is not proportional to the value for the concentration of sulphuric acid in aerosol drops mentioned earlier.

Keeping in mind that at the present time high-altitude planes are able to reach the aerosol layer in the stratosphere with loads of approximately fifteen tonnes we find that such a massive reagent may be delivered into lower stratosphere by several regularly scheduled aircraft equipped with devices for burning sulphur in the atmosphere.

It should also be noted that the volume of reagents thus contributed to the atmosphere for influencing the climate is many thousands of times less than is the quantity of matter that enters the atmosphere under the influence of human activity, and which has been estimated at hundreds of millions of tonnes per year. Similarly, the quantity of reagents that eventually is deposited on the Earth's surface is altogether negligible. For the rates of expenditure that have been cited it would be equal to 0.2 mg of sulphur per square meter per year on the average, which is approximately one-thousandth of the natural contribution of sulphur through atmospheric precipitation. It is clear that such small quantities are of no consequence from the point of view of environmental pollution.

It should be noted that while the calculations on which these conclusions are based are highly approximative in nature, there are grounds for believing that further improvements in their precision in later studies will not change the conclusion concerning the fundamental possibility of influencing climate through modern technological equipment by regulating the concentration of aerosols in the stratosphere.

Let us note that the selection of a region for contributing the reagents in such a project for influencing the climate is not very important, since, as has been established by data relating to the dissemination of products of volcanic eruptions, reagents from any point within extratropical zones are quickly disseminated throughout the entire hemisphere. At the same time the study of circulation in the lower stratosphere may make it possible to facilitate a selection of optimal regions and periods of time for disseminating reagents in a manner that will provide for their most effective utilization.

The method for influencing the climate that has just been considered is intended to prevent or else weaken climatic changes that may arise in the next few decades under the influence of economic activity. Such influences, however, do not exhaust the possibilities of modern technology. It would appear that in the near future it will be technically possible to influence climate in ways that will cause the changes of an order of magnitude larger than the fluctuations in climate that took place in the 1920s and 1930s, i.e., to produce declines in global temperature by several degrees.

Influences of such magnitudes may be necessary in the twenty-first century, when as a result a major increase in the production of energy the temperature of lower layers of the atmosphere may increase substantially. Reductions in the transparency of the stratosphere under such conditions may prevent undesirable climatic changes.

Let us note that these types of influence may apparently be carried out in the absence of forecasts of natural changes in climate, by relying on current information about the concentration of aerosols in the stratosphere, which is received from data gathered by high-altitude monitoring planes or else from data derived from ground actinometric observations on the Earth's surface. Since average concentrations of aerosols in the stratosphere under natural conditions change relatively slowly it is possible to regulate the scale of activities intended to maintain the aerosol concentration at an optimal level by interrupting such activities when it attains the required magnitude under the influence of natural processes.

Let us now consider the possibilities of climatic changes resulting from major increases in the number of flights of aircraft in the stratosphere. This problem has been considered in a collective monograph prepared by American scientists (SCEP, 1970), in which it is noted that in 1985-1990 civil aviation will employ approximately 500 aircraft flying at heights of 18-20 kilometers. The by-products of the burning of fuels by these aircraft will contribute 63,000 tonnes of sulphur dioxide each year to the lower stratosphere, which corresponds to a volume of sulphate particles of approximately 100,000 tonnes.

In considering this question it is important to take into consideration the fact noted earlier that prolonged changes in radiation exert much greater influence on the atmospheric thermal regime than similar changes over a short period of time.

Assuming that the major part of the stratospheric flights will take place in the Northern Hemisphere the magnitude of yearly increases in the volume of sulphate particles in the stratosphere may be taken as 70,000-80,000 tonnes and the overall mass of aerosols as not less than 200,000 tonnes.

For an average life-time of particles in the aerosol layer of two years stratospheric flights will increase the concentration of these particles in the Northern Hemisphere by approximately 400,000 tonnes. This represents more than one-half of the quantity of aerosols that may produce the substantial changes in climate mentioned earlier. Apparently that quantity should not be viewed as insignificant.

It is worthwhile noting the possibility of influencing the climate through changes in sulphur content in the fuels employed by stratospheric aircraft. The calculation that has been cited is based on the assumption that the fuel employed in stratospheric flights will have a relatively low sulphur content of approximately 0.05 per cent (SCEP, 1970). The authors of those calculations also consider the possibility that fuels with a sulphur content of 0.3 per cent might be used as well as fuel in which the sulphur content is reduced to 0.01 per cent, even though the last type of fuel will apparently be more expensive than fuel with a la ge content of sulphur.

It follows from the data that have been cited that a change in the quantity of sulphur in fuels from 0.01 to 0.30 per cent produces changes in the mass of aerosol particles in the Northern Hemisphere's stratosphere from a figure of 80,000 tonnes to 2,400,000 tonnes. While the first of these figures does not produce substantial changes in climate, the second exceeds changes in the quantity of aerosols that are needed for solving the tasks of climate regulation. In this connection there arises the prospect of influencing the climate without additional expenditures through changing types of fuel employed by stratospheric aircraft. In order to increase the concentration of aerosol particles when actual levels are too low fuels with higher sulphur content should be used, while to reduce their concentration low sulphur-content fuels are required.

Thus it is possible to view the regulation of air traffic in the stratosphere as one of the elements of a system of global climate regulation. That problem calls for further studies in connection with the need to clarify the effect of stratospheric flights on other elements of the meteorological regime, including the stratosphere ozone content. More generally it should be expected that the increased numbers of flights in the stratosphere may lead to changes in climatic conditions whose nature should be anticipated.

It follows from the data that have been cited that relatively modest changes in the balance of stratospheric aerosols may substantially influence air temperature at the Earth's surface.

The possibility of applying such a method for preventing natural fluctuations in climate that reduce the intensity of water exchange in regions of insufficient moisture is also a matter of interest.

In order to solve these problems further advances in the theory of climate are of major importance, since currently existing simplified theories cannot identify all possible changes in the meteorological regime of different regions of the globe that result from influences on the stratospheric aerosol layer.

At the same time a further development of numerical models is needed, encompassing the entire range of processes that determine the state of the global ecological system, in order to calculate changes that may follow within various elements of the biosphere as a result of changes in climate.

The international aspect of the problem of anthropogenic changes in climate should be emphasized. This becomes especially important in developing large-scale activities to influence the climate, since influences on global climate may bring about alterations in natural conditions on the territories of many countries, and since the nature of these changes may differ in different regions.

Conclusion

Global ecology studies the biosphere as a whole, i.e., the ecological system that encompasses the entire globe. That system is regulated by relations between living organisms and their environment. Mutual interactions between organisms and the atmosphere, which disseminates changes in the environment produced by organisms to the entire planet, are particularly important for global ecological studies. The close dependence of living nature on the atmosphere is seen from the fact that to a large extent the atmosphere is created by living organisms, and in its turn exerts an influence on them that in many cases determines their activities.

In many respects the global ecological system is integrated, and this makes it possible to study its evolution as that of an integrated structural formation. In understanding the history of the global ecological system much importance attaches to the problem of its stability. In studying that problem, as well as many other problems in the evolution of the global system, it is necessary to rely on numerical models based on an analysis of cause-and-effect relations among that system's components.

Initial experiments in devising such models lead to the conclusion that the stability of the global ecological system in relation to the influence of external factors is not very great. It is possible that a partial disruption of that stability has occurred in critical ages of the Earth's geological history, when they produced changes in prevailing types of fauna. It is also possible that the prolonged existence of the global ecological system is a result of a statistically relatively improbable combination of circumstances that are not likely to occur frequently in the Universe.

The conclusion that the stability of the global ecological system is not very great is confirmed by the enormous consequences of man's influence on that system over a period of time that is very short from the point of view of the history of life on our planet generally. Long before the beginning of the age of industrial development man altered substantially the plant cover and animal life over a large part of the surface of continents. In recent centuries, and especially in recent decades, man's influence on the environment has sharply increased. Since these activities have frequently taken place without consideration for possible ecological consequences their outcome has often produced a deterioration in the environment.

In recent times human activities have not only greatly intensified influences on the environment, but have also begun to influence large-scale atmospheric processes. This makes possible changes in all components of the global ecological system.

While in many cases contemporary human activities have caused substantial harm to the environment, changes in natural conditions produced by man are not as a rule irreversible, and in such a context it may be said that at the present time there does not exist a global ecological crisis. Nevertheless if present rates of industrial development continue such a crisis may arise in the relatively near future.

Keeping in mind the finding that the stability of the global ecological system is limited, it may be expected that in the absence of ecological planning further technological progress will lead to specific disruptions in that system's stability.

Accordingly, the problem of control over anthropogenic changes in the environment is becoming very relevant. The need for forecasting possible changes in the natural environment and for devising methods of influencing the global ecological system that may prevent its development in undesired directions is evident.

The tasks relating to the preservation of natural resources and to comprehensive studies of the influence of large-scale production activities on the development of natural processes have been stated clearly at the 25th Congress of the CPSU. In a document entitled "Guidelines for the Development of the National Economy of the USSR for 1976-1980" it is stated: "To further develop scientific principles for the rational use and conservation of soils, of mineral wealth, flora and fauna, air and water basins. To do wider integrated World Ocean studies. To continue work on weather forecasting and that of natural disasters." (*Documents and Resolutions. XXVth Congress of the CPSU*, Moscow, 1976, p. 233.)

Detailed studies of global ecological processes are needed to solve these problems. These are still at an early stage of development. It is especially important to construct numerical models of global ecological systems that should serve as a basis for carrying out these tasks.

In order to prevent changes in global ecological processes that are unfavourable to man effective international forms of cooperation are also needed. The global ecological system is the property of all of mankind. This property must be preserved from irresponsible activities that may produce harm to the environment within which man exists.

This is why in 1974 the Soviet Government addressed to the United Nations a proposal to sign a multilateral agreement prohibiting the manipulation of the environment and of the climate for hostile objectives. That proposal was approved by the UN General Assembly, which adopted a decision to prepare a draft for such an Agreement.

In 1976, at the 31st Session of the UN General Assembly a convention was approved on the prohibition of military or any other hostile utilization of means for influencing the environment. The assembly recommended to all states that they sign and ratify the Convention.

This may be viewed as the beginning of comprehensive international measures that will be carried out in order to preserve and improve our planet's natural conditions.

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